QUARTERLY JOURNAL

OF THE

ROYAL METEOROLOGICAL SOCIETY

Editor: J. S. SAWYER

Vol. 85 APRIL 1959 No. 364

CONTENTS

PAGE

A. M. OBUKHOV and A. M. YAGLOM—On the microstructure of atmospheric turbulence—a review of recent work in the U.S.S.R. 81

H. ISRAËL—Atmospheric electrical agitation 91

R. H. DOUGLAS and W. HITSCHFELD—Patterns of hailstorms in Alberta 105

D. G. JAMES—Observations from aircraft of temperatures and humidities near strato-cumulus clouds 120

R. S. SCORER and H. KLIEFORTH—Theory of mountain waves of large amplitude .. 131

BETSY WOODWARD—The motion in and around isolated thermals 144

F. H. W. GREEN—Some observations of potential evaporation, 1955-57 152

R. G. VINES—Wind stress on a water surface: measurements at low wind speeds with the aid of surface films 159

Shorter Contributions—

B. C. V. ODDIE—The composition of precipitation at Lerwick, Shetland 163

R. A. HAMILTON—The pressure gradient at the edge of an ice sheet 166

R. A. HAMILTON—The diurnal variation of pressure on the Greenland ice sheet .. 168

Discussions—

R. C. RAINEY—Some observations on flying locusts and atmospheric turbulence in Eastern Africa (*Q.J.*, **84**, p. 334) 171

C. G. JOHNSON—Vertical distribution of aphids in the air and the temperature lapse rate (*Q.J.*, **83**, p. 194) 173

R. A. HAMILTON—The meteorology of North Greenland during midsummer and midwinter (*Q.J.*, **84**, p. 142 and p. 355) 176

Proceedings at meetings of the Society 179

Reviews—

Polar atmosphere symposium, Pt. 1, Meteorology section, Ed. R. C. Sutcliffe (H.H.L.); *The upper atmosphere*, H. S. W. Massey and R. F. L. Boyd (R.J.M.); *Rockets and satellites*, Ed. L. V. Berkner (D.H.M.); *Everyday meteorology*, A. Austin Miller and M. Parry (A.H.G.); *Wetter- und Meereskunde für Seefahrer*, J. Krauss and W. Stein (H.H.L.); *Combination of observations*, W. M. Smart (N.G.) 182

Obituary Notice 186

PUBLISHED AND SOLD BY THE
ROYAL METEOROLOGICAL SOCIETY
49 CROMWELL ROAD, LONDON, S.W.7
PRICE THIRTY SHILLINGS ($4.50)
ANNUAL SUBSCRIPTION £5 - 4 - 0 ($16), POST FREE

ROYAL METEOROLOGICAL SOCIETY

Founded on 3 April 1850, for the advancement of meteorological science. The Society was incorporated by Royal Charter in 1866.

MEMBERSHIP

Membership is open to all persons genuinely interested in meteorology. The annual subscription is as follows :

Fellows	£4-4-0 (Entrance fee, 1 guinea)	
Foreign Members ..	£4-4-0	„ „
Associates	£3-0-0	„ „
Students	£1-10-0	

All Members receive free the regular publications of the Society, including the monthly publication *Weather* (annual subscription to non-members 24s. plus 2s. 6d. postage).

The Descriptive Account of the Society and an application form will be sent on request to the Assistant Secretary.

Quarterly Journal

EDITING COMMITTEE

J. S. Sawyer — *Editor and Chairman*

G. A. Corby A. G. Forsdyke R. F. Jones F. Pasquill

H. L. Penman G. D. Robinson R. S. Scorer

The primary purpose of the *Quarterly Journal* is to publish results of original research and reports of the Society's scientific meetings. Papers intended for the *Journal* **should be submitted in accordance with standards laid down by the Council in the Society's memorandum** ' Preparation of Papers for the *Quarterly Journal* ' which is available on request. Failure to conform with these standards involves the Society's officers and staff in much needless work and the Society in unnecessary expense. **Manuscripts will therefore not be considered for publication unless they conform reasonably closely with the procedure laid down.** The attention of authors is drawn particularly to the requirements in regard to mathematics and diagrams and the Assistant Secretary will be pleased to advise authors on their special problems of illustration before papers are formally submitted. Diagrams should, if possible, be submitted flat, not rolled, and a duplicate rough set for refereeing purposes should be included. Where rulings on a diagram are desirable the closest acceptable spacing is approximately 1 cm and this may be done on **blue-lined** graph paper by thickening the centimetre rulings. Manuscripts may be submitted in duplicate to expedite refereeing, but authors must retain complete copies in their possession.

Price to non-members is £5-4-0 (post free) per annum and £1-10-0 for single issues (postage extra).

PUBLICATIONS

Copies of the publications *Centenary Proceedings of the Royal Meteorological Society*, 1950 (£1. 1s.) and *Proceedings Toronto Meteorological Conference* 1953 (£1. 10s.) are still available.

Reprints of papers recently published in the *Quarterly Journal* are generally available, the prices being as follows (postage extra): 8 pp., 1s. 2d.; 16 pp., 1s. 10d.; 20 pp., 2s. 2d.

Weather Lore. Fourth edition of the late Richard Inwards's unique compilation, revised, augmented and edited for the Society by E. L. Hawke. Published in 1950. Price 15s. Price to Members, 10s. 6d., postage 9d.

Rainfall Atlas of the British Isles. Published in 1926. Price to Members, 15s. post free.

Collected Scientific Papers of William Henry Dines, F.R.S. This memorial volume contains reproductions of nearly sixty papers and other original communications by W. H. Dines. Price 15s., postage 1s.

Cloud Photographs. A series of 24 post-cards, reproduced from the well-known Clarke Collection, illustrating high, medium and low types of cloud. Price 7s. 6d., complete set, postage 6d.

Cloud study : a pictorial guide. By F. H. Ludlam and R. S. Scorer. 75 photographs. Published in 1957 under the auspices of the Society by John Murray. 12s. 6d. (11s. to Members), postage 1s.

THE SOCIETY'S OFFICES

Assistant Secretary — R. H. Levinson, M.A., A.C.A.

49 Cromwell Road, London, S.W.7.

(*Telephone :* KENsington 0730)

The offices and library of the Society are open daily (Monday to Friday) from 10 a.m. to 5 p.m. All enquiries and correspondence concerning the Society and its publications should be addressed to the Assistant Secretary.

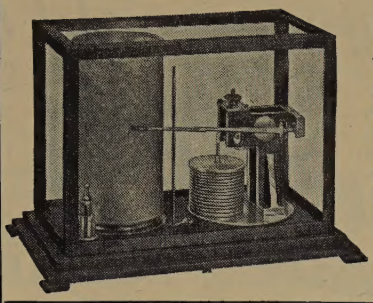
SHORT & MASON

Manufacturers

of the finest

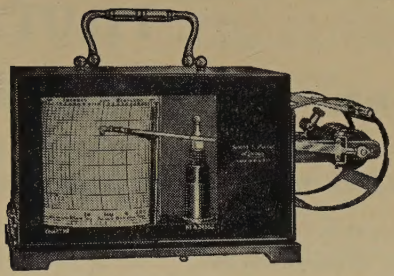
Scientific

Instruments



MICRO-BAROGRAPH

This recorder has a very open scale in which 2½ linear inches equal 1" barometric pressure. Essential for weather forecasting, barometric pressure studies and survey work. Supplied with 24-hour or 7-day revolution clock and charts in inches, millimetres and millibars. Prompt delivery.



THERMOGRAPH

This is a highly sensitive bimetallic coil instrument having a low-drum weekly clock, exposed element and "gate" suspension pen arm. Supplied complete with 6" setting Thermometer, a year's supply of charts and ink. Prompt delivery.

METEOROLOGICAL THERMOMETERS

Standard Maximum, Minimum, Ordinary and 'Solar Radiation' Thermometers can be supplied either sheathed or mounted. These thermometers can be supplied with N.P.L. Certificates if required. Prompt delivery.

ALSO BRASS CASE BAROMETERS · SURVEYING ANEROID BAROMETERS · MERCURIAL BAROMETERS · HYGROMETERS HYGROGRAPHS · STANDARD THERMOMETERS · WIND SPEED AND DIRECTION INDICATORS · ANEMOMETERS RAIN GAUGES · GAS MEASURING INSTRUMENTS, etc., etc.

Catalogue and Price List sent on request

SHORT & MASON LTD. ANEROID WORKS, 280, WOOD STREET, WALTHAMSTOW, LONDON, E.17. TELEPHONE: COPPERMILL 2203/4.

Why you should specify

Beritex REGD SEAMLESS

SOUNDING AND PILOT BALLOONS

Made by **THE GUIDE BRIDGE RUBBER CO. LTD**

They're the finest available for meteorological work because:—

Beritex Balloons have years of research and development behind them.

Every balloon is guaranteed and has been rigorously tested and individually inspected.

Made from the finest Rubber Latex they require no special pre-flight treatment.

* * *

Les ballons 'Beritex' ont des années de recherches et de mise au point derrière eux.

Chaque ballon est garanti et a fait l'objet d'un essai rigoureux et d'une vérification individuelle.

Fabriqués en latex de caoutchouc de la meilleure qualité ils n'ont besoin d'aucun traitement spécial avant l'envol.

* * *

Los globos Beritex están respaldados por muchos años de investigación y fabricación.

Todos los globos están garantizados y han sido rigurosamente probados e individualmente verificados.

Los globos, fabricados con inmejorable caucho látex, no requieren ningún tratamiento especial antes del vuelo.

OVERSEAS SALES

Phillips Patents Limited · 250 Western Avenue · Acton · London · W3

HOME SALES

Guide Bridge Rubber Co. Ltd · Vulcan Mill · Butcher Lane · Bury · Lancs.

Important New Books from Pergamon Press

International Series on Aeronautical Sciences and Space Flight

VISTAS IN ASTRONAUTICS, Vol. II

Editors : MORTON ALPERIN *and* H. F. GREGORY, *Air Force Office of Scientific Research, U.S.A.*

Contains papers presented at the Second Astronautics Symposium by scientists, research personnel from industry and the military. These papers cover the entire field of original research, both theoretical and experimental, and deal with such subjects as vacuum chamber research, cosmic debris of interplanetary space, instrumentation of an Earth-Mars flight, space navigation, launching of space vehicles, the comparative performance of ballistic and low thrust vehicles, exploration of the moon and the lunar dust level. None of the material has been published before.

330 pages April Price 105s. net ($15)

ADVANCES IN AERONAUTICAL SCIENCE (2 Vols.)

Editors : Th. VON KÁRMÁN, A. M. BALLANTYNE, H. BLENK, R. R. DEXTER, M. ROY, H. L. DRYDEN

In these two volumes will be published the 46 papers presented by the world's outstanding specialists in the field of aeronautics and astronautics at the First International Congress of the Aeronautical Sciences held in Madrid, Sept. 1958. The meeting, under the leadership of Professor Theodore von Kármán, was attended by some 600 delegates from 25 countries, including an official delegation from the U.S.S.R.

1,200 pages May Price £8 net ($25)

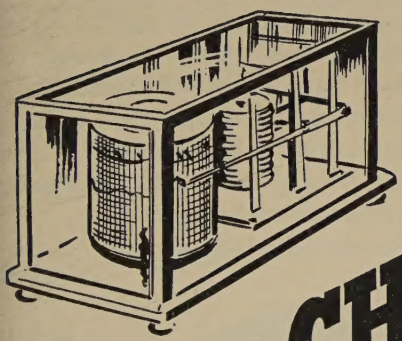
COLLECTION OF THEORETICAL PAPERS IN AERODYNAMICS

Issued by the Central Aerohydrodynamics Institute of the U.S.S.R. In preparation

Please write for fully descriptive leaflets

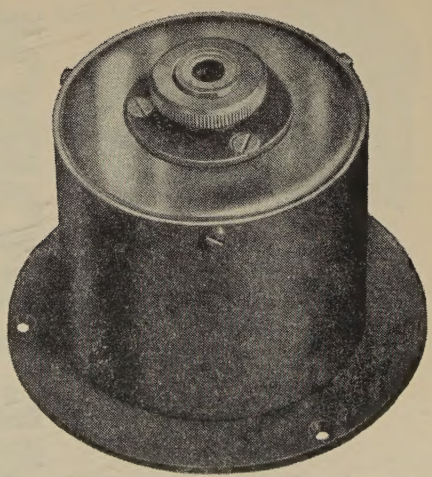
PERGAMON PRESS *LONDON · NEW YORK · PARIS · LOS ANGELES*

4 & 5 Fitzroy Square, London, W.1. 122 East 55th Street, New York 22, N.Y.
24 Rue des Ecoles, Paris, Ve



HORSTMANN CHART CLOCK

NEWBRIDGE WORKS BATH REGISTERED TRADE MARK



Horstmann Chart Clocks are used extensively by the Air Ministry Meteorological Office. Two types are available, one to drive chart drums revolving once in every 24 hours and the other for 7 day rotation. The clockwork movements of both types have a centre spindle drive.

The movements are of robust construction and incorporate very high quality jewelled platform escapements, which ensure extreme accuracy in time keeping.

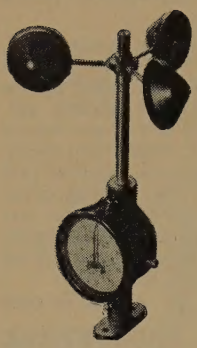
Chart Clocks having speeds other than those mentioned above can also be supplied to special order.

THE HORSTMANN GEAR CO. LTD. · NEWBRIDGE WORKS · BATH · Tel.: Bath 7241

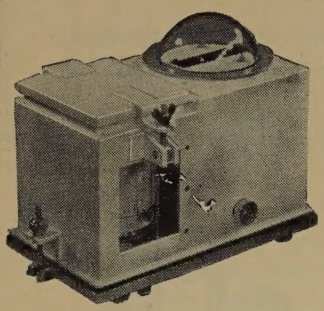
METEOROLOGICAL INSTRUMENTS

for measuring

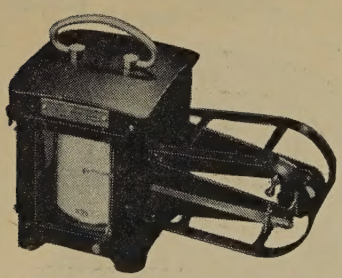
- ★ WIND
- ★ PRESSURE
- ★ TEMPERATURE
- ★ HUMIDITY
- ★ PRECIPITATION
- ★ EVAPORATION
- ★ RAINFALL
- ★ SUNSHINE
- ★ ETC., ETC.



PORTABLE ANEMOMETER
Has a range of 25/5,000 feet per minute.



ACTINOGRAPH
Designed by the British Meteorological Office.



THE THERMOGRAPH
Incorporates bimetallic coil which gives rapid response to temperature changes.

Whenever meteorological phenomena have to be recorded, Casella can provide instruments that will give you the most reliable readings. Whether they are well tried standard patterns or specially designed instruments for highly specialised requirements, all embody the high standards of workmanship and finish that have been the Casella manufacturing policy for over 140 years. *The new Casella catalogue of meteorological instruments is now ready. Send for your copy.*

CASELLA
LONDON

C. F. CASELLA & CO. LTD.
Regent House, Fitzroy Square,
London, W.1
Telephone EUSton 3944

Members Scientific Instrument Mauufacturers' Association of Great Britain and the British Scientific Instrument Research Association.

QUARTERLY JOURNAL
OF THE
ROYAL METEOROLOGICAL SOCIETY

Vol. 85 APRIL 1959 No. 364

551.551.551.511.6

On the microstructure of atmospheric turbulence – a review of recent work in the U.S.S.R.

By A. M. OBUKHOV and A. M. YAGLOM
Institute for Physics of the Atmosphere, Moscow

(Communicated by P. A. Sheppard. Manuscript received 19 September 1958)

SUMMARY

Owing to turbulence the instantaneous distribution of wind velocity, temperature, humidity and other meteorological elements is of a very complicated character, the elements at different points undergoing random and often very considerable fluctuations. Similarly, the values of the same elements at a fixed point and at different moments are characterized by random fluctuations having a very wide spectrum. Space and time fluctuations of meteorological elements are of general geophysical interest since they characterize atmospheric turbulence which is one of the most important features of atmospheric motion. Moreover, the study of fluctuations has an applied significance as one needs to know the statistical characteristics of fluctuations for the determination of the rate of the turbulent exchange in the surface layer, for the theory of meteorological instruments and for a number of problems of atmospheric optics, atmospheric acoustics and radiometeorology.

Because of their wide spectrum (both in space and time) the methods of study of these fluctuations vary according to the particular part of the spectrum which we are interested in. Here we shall mainly deal with the problem of small-scale turbulence (of the micro-inhomogeneities of the atmospheric fields) which are characterized by periods of a few hundredths of a second to some seconds and space scales of a few centimetres to some metres. Our considerations will mostly refer to the surface layer which has been most studied. In the U.S.S.R. a number of theoretical and experimental investigations concerning the microstructure of the principle meteorological fields, mainly those of the wind velocity and temperature, have recently been carried out. The results of these investigations are in satisfactory agreement. Simultaneously some special problems of the physics of the atmosphere using the above-mentioned results have been treated. A brief review of all the investigations is given below.

1. Statistical characteristics of turbulence

We shall begin with the problem of what particular characteristics of the meteorological fields are best fitted for the description of their microstructure (to avoid ambiguity we shall speak here only about the space micro-structure). In the mathematical theory of random functions the correlation function

$$B(r) = \overline{u(M)\,u(M')}, \quad r = |\overrightarrow{MM'}| \quad . \quad . \quad . \quad (1)$$

is ordinarily used as a principal statistical characteristic of the random function $u(M)$ at point M (the bar will always denote averaging). It should be noted that the role of statistical characteristics of this kind in turbulence theory was first indicated by Keller and Friedmann (1924) at the first International Congress on Applied Mechanics in Delft. For meteorology, however, the structure function

$$D(r) = \overline{[u(M) - u(M')]^2} \quad . \quad . \quad . \quad . \quad (2)$$

introduced by Kolmogorov (1941 a, b) is more suitable. In the case of vector fields such as the field of wind velocity $\vec{u}(M)$ we have to speak about two structure functions : the longitudinal function $D_{ee}(r)$ and the lateral function $D_{nn}(r)$:

$$D_{ee}(r) = \overline{[u_e(M) - u_e(M')]^2}, \quad D_{nn}(r) = \overline{[u_n(M) - u_n(M')]^2}, \quad . \quad (3)$$

where $u_e(M)$ and $u_n(M)$ are components of the vector $\vec{u}(M)$ parallel and normal to the

vector separation $\vec{r} = \overrightarrow{MM'}$. It is understood that if the correlation function depends upon the distance r between M and M' alone, then the structure function will depend on the same distance and

$$D(r) = 2[B(0) - B(r)]. \quad . \quad . \quad . \quad . \quad (4)$$

The use of either of the two functions in this case appears to be equally justified. One should keep in mind, however, that for hydrodynamical fields, the structure function depends only upon the eddies the scales of which are less than, or comparable with, r, while all the eddies, even the biggest, contribute considerably to the correlation function. As the biggest eddies are usually not isotropic, the function B cannot be dependent on the distance r alone, whereas for the function D, if r is not very great, such an assumption is completely justifiable. The well-known Kolmogorov theory of locally isotropic turbulence is based on these considerations. This theory initiated the investigation of the microstructure of meteorological fields and remains a guiding clue in the investigations. For atmospheric turbulence, where the mean values of the meteorological elements are generally unsteady, the advantage of the structure functions is particularly obvious. When considering difference-values of the meteorological elements we need not fear the influence of ' the evolution of the mean-level.' In this connexion the structure functions turn out to be less subject to fluctuations, i.e., they can be correctly determined with considerably less averaging. Moreover, the computation for the determination of the structure function proves to be simpler than that for the correlation function (1). For these reasons the structure functions are preferably used as the main statistical characteristics in the study of meteorological fields. They are used even when, for one reason or another, we need to know the correlation function in particular; it is more convenient to determine the function $D(r)$ and then to find the fluctuation intensity $\overline{u^2} = B(0)$ in addition and to use the formula (4)*.

2. Shapes of structure functions

Theoretical considerations and experimental evidence both show that the shapes of the structure functions of meteorological fields under different conditions of measurement have, as a rule, some common features. A typical shape of such a function is shown in Fig. 1. Near the origin this function is a small parabolic curve i.e., $D(r) \simeq ar^2$ in this range. The presence of this range is determined by dissipation processes (by viscosity in the case of the velocity field, heat conductivity in the case of the temperature field and molecular diffusion in the case of the humidity field). Its size is characterized by an internal (local) scale of atmospheric turbulence l of order 1 cm in the surface layer of the atmosphere. There follows a considerable range where the structure function is well described by a simple power function with some power index α which is less than 2

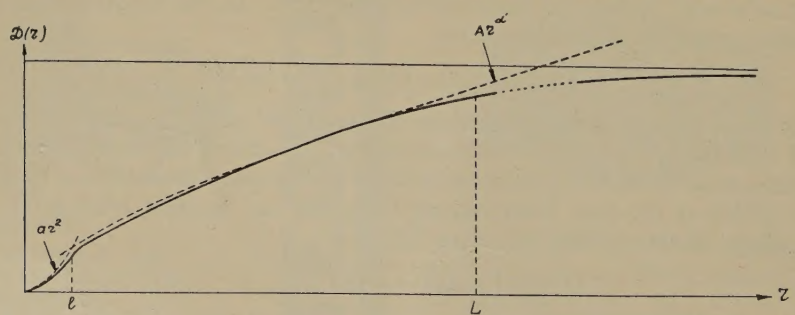


Figure 1. A typical shape of a structure function.

* Since the correlation function $B(r)$ in all actual cases dies away as $r \to \infty$, then theoretically the constant $B(0)$ can be found from the asymptote of the structure function $D(r)$ as $r \to \infty$ (for according to (3) $D(r) \to 2B(0)$ if $r \to \infty$). In practice, however, for meteorological fields, the approach of $D(r)$ to $2B(0)$ is very slow, so that the determination of the fluctuation intensity according to the measurements of he structure function is found to be unreliable.

(equilibrium range). Finally, in the range of very large scales the growth of the function $D(r)$ slows up and when the functions are statistically steady the function approaches the constant (equal to $2B(0)$)*. The distance L beyond which the growth of the function $D(r)$ is negligibly small, is called an external scale of turbulence. This scale is determined by the sizes of the biggest eddies existing in the atmosphere, i.e., it is determined by the geometry of the flow and the external sources of energy. In the surface layer it has the same order as the distance from the ground (if the points M and M' are on the same level it is usually somewhat larger than the distance from the ground) and in the free atmosphere it often has the order of many hundred kilometres. We shall not deal below with the range of scales exceeding the external scale of turbulence. This scale will not play any role in the geophysical problems treated in the later part of this paper, so that the equilibrium range can be considered here to expand to infinity.

3. Theoretical results concerning structure functions

We shall point to the principal theoretical results concerning the structure of the meteorological fields in the equilibrium range $l < r < L$ and in the 'dissipation range' $r < l$. For the field of wind velocity (the field of velocities of locally isotropic turbulent flow) the main result was obtained by Kolmogorov (1941 a, b). He showed theoretically that the power index α here must equal $\frac{2}{3}$. In the same year the identical result was obtained by Obukhov (1941) from considerations of the distribution of energy of turbulent flow in wave-number space. So, if $l < r < L$ equalities such as

$$D_{ee}(r) = A_1 r^{\frac{2}{3}} \text{ and } D_{nn}(r) = A_2 r^{\frac{2}{3}} \quad . \quad . \quad . \quad (5)$$

must hold. From the equation of continuity it is easy to deduce that the coefficients A_1 and A_2 are connected by $A_2 = \frac{4}{3} A_1$. The absolute value of these coefficients is dependent first of all on a mean energy dissipation ϵ; the value ϵ can be rather exactly evaluated from observations of the vertical distribution of the mean wind velocity†. If $r < l$ it is obvious that

$$D_{ee}(r) = a_1 r^2 \text{ and } D_{nn}(r) = a_2 r^2 \quad . \quad . \quad . \quad (6)$$

where $a_2 = 2a_1$, while the coefficients a_2 and a_1 are simply expressed in terms of ϵ and the viscosity, ν.

If we apply considerations, similar to those used by Kolmogorov, to the pressure field $p(M)$ we shall come to the following result

$$\frac{1}{\rho^2}\overline{[p(M) - p(M')]^2} = \frac{1}{\rho^2} D_{pp}(r) = A_3 r^{\frac{4}{3}}, \quad l < r < L \quad . \quad . \quad (7)$$

(Obukhov 1949). It is interesting to note that the theoretical evaluation enables us to foresee that the coefficients A_3 will coincide with A_1^2 to within 10 to 15 per cent (so that $\frac{1}{\rho^2} D_{pp}(r) \simeq [D_{ee}(r)]^2$). It is somewhat more complicated to evaluate the coefficient a_3 in the formula

$$\frac{1}{\rho^2} D_{pp}(r) = a_3 r^2, \quad r < l. \quad . \quad . \quad . \quad . \quad (8)$$

This coefficient characterizes the mean square of local gradients of pressure (see Heisenberg 1948, Yaglom 1949b, Obukhov and Yaglom 1951). We shall indicate here that the mean-squared acceleration $\left[\overline{\left(\frac{d\vec{u}}{dt}\right)^2}\right]^{\frac{1}{2}} \approx 1{\cdot}7\, \nu^{-\frac{1}{4}} \epsilon^{\frac{3}{4}}$ determined from these evaluations turns out to be rather considerable. Thus Yaglom (1949) and Obukhov and Yaglom (1951) show that such an acceleration at a height at 1·5 m, when the roughness length is 3 cm, and

* In practice the approach to the constant is sometimes accompanied by smooth pulsations, so that for certain intervals the function $D(r)$ even turns out to be decreasing.

† From Kolmogorov's theory it follows that $A_1 = C_1 \epsilon^{\frac{2}{3}}$ and $A_2 = C_2 \epsilon^{\frac{2}{3}}$, where C_1 and C_2 are constants of order unity and $C_2 = \frac{4}{3} C_1$. The value ϵ in the surface layer, at the height z is equal to V_*^3/kz, where $k \approx 0{\cdot}4$ and V_* is a friction velocity, determined from the profile of the mean velocity.

the mean wind velocity V is about 5 m sec^{-1}, reaches the acceleration of gravity, g. If V is further increased the acceleration will increase rapidly, proportional to V^2.

Let us come to the discussion of the microstructure of the temperature. The present authors first applied the theory of locally isotropic turbulence to the study of this microstructure in 1949 (1949a; 1949b). Later on, Corrsin (1951) using the spectral approach to the problem came to the same result*. The entropy equation (taken instead of the energy equation, used in the theory of microstructure of the wind field) plays a principle role in these considerations. In the range of middle scales, redistribution of entropy takes place owing to turbulent mixing. The increase of the entropy due to thermal conductivity occurs in the range of small scales (dissipation range). If we assume that the motion in the range of small and middle scales is approximately steady then we can obtain the following result

$$\overline{[T(M) - T(M')]^2} = D_{TT}(r) = A_4 r^{\frac{2}{3}}, \quad l_1 < r < L \quad . \quad . \quad (9)$$

The coefficient A_4 here is determined mainly by the mean dissipation of energy ϵ and the mean rate of increase of entropy $N = \kappa \overline{(\text{grad } T)^2}$, where κ is the coefficient of the thermometric conductivity†; the value N for the temperature field is an analog of the dissipation energy ϵ. The scale l_1, coincides in order with the local scale l for the field of velocity. This circumstance is connected with the fact that the coefficients of viscosity and heat conductivity are of the same order‡. If $r < l_1$ it follows that

$$D_{TT}(r) = a_4 r^2 \quad . \quad . \quad . \quad . \quad (10)$$

where the coefficient a_4 is simply expressed through N and κ.

The analysis of the microstructure of the temperature field in the atmosphere is also applicable to the structure of the concentration field of any passive substance; the coefficient of the thermometric conductivity κ being replaced by the coefficient of molecular diffusion, as in the study of the microstructure of the humidity field.

4. Equipment for micropulsation measurements

The experimental measurement of micropulsations of the meteorological fields requires the construction of special equipment with a very small time lag, of the order of 0·01 sec, high sensitivity and small size of sensing elements (to avoid considerable averaging in space). In the case of the field of the wind velocity it is obvious that the most suitable principle for the construction of such equipment is that of the hot-wire anemometer. A very sensitive hot-wire anemometer satisfying all the indicated conditions was constructed and applied to measurements in the atmosphere as early as 1936 by the German scientist Gödecke (1936). In the Soviet Union the use of a hot-wire anemometer for the experimental investigation of the laws of the local structure of turbulence was begun by Obukhov (1942) just after the appearance of the corresponding theoretical results in 1941, but it was interrupted by the war and renewed only in 1946-47 with the participation of Krechmer who was also engaged in the construction of a special instrument (microthermometer) for measuring and recording micropulsations of the temperature. These instruments had acquired their final form only by 1950. Now we shall in brief remind you of the principles of their action. (They are described in detail by Krechmer (1954)).

Two hot-wire anemometers with platinum wires of 20 μ diameter and 2 cm long were used simultaneously. These wires serve as sensing elements of the equipment. The difference in resistance of these two wires is measured with a Wheatstone bridge. Placing the wires at two different points of space we can determine the difference between the wind velocity at these points. The same equipment can be used for measuring the

* Inoue's work (1950) on the same theme had appeared a little earlier than Corrsin's. In this work for $D_{TT}(r)$ the law of the form (9) but with a power index $\frac{4}{3}$ instead of $\frac{2}{3}$ had been deduced. However, this conclusion was erroneous as was proved, in particular, when analysing the empirical material on the microstructure of the temperature field in the atmosphere obtained separately in the U.S.S.R. (Krechmer 1952; Tatarskiy 1956a) and in Japan (Shiotani 1955).

† From the theory it follows that $A_4 = C_4 N \epsilon^{-\frac{1}{3}}$, where C_4 is a constant of order unity.

‡ As it is easy to show that $l \sim \nu^{\frac{3}{4}} \epsilon^{-\frac{1}{4}}$ and $l_1 \sim \kappa^{\frac{3}{4}} \epsilon^{-\frac{1}{4}}$.

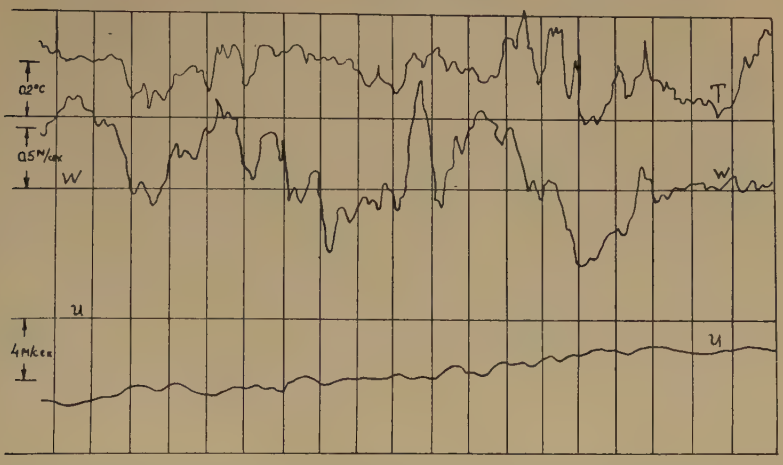


Figure 2. The pattern of photo-records of microfluctuations of the instantaneous values of the horizontal component of the wind velocity u, the vertical component of the velocity w and temperature T. (The vertical traces are time marks, appearing every 0·1 sec).

vertical component of the velocity. For this purpose two sensing wires are placed to be $\pm 45°$ to the horizontal plane. The time lag of this equipment is about 0·01 sec and its sensitivity is about 1 cm sec^{-1}.

The microthermometer uses the same platinum wires of 20 μ diameter and some centimetres long as sensitive elements, but in this case they are used as resistance thermometers. For this purpose, we pass a very feeble alternating current of 4000 c/s through the wires so that there is practically no heating. Further, the current is amplified with the application of the phase-detector. The time lag of the instrument is of the same order as for the hot-wire anemometer and the sensitivity is of the order of 0·01°C.

It should be noted that almost simultaneously a very similar equipment was constructed by Shiotani (1950, 1955) in Japan, Later on, equipment with small time lag for measuring micropulsations of the meteorological fields was also constructed by MacCready (1953) in the U.S.A.

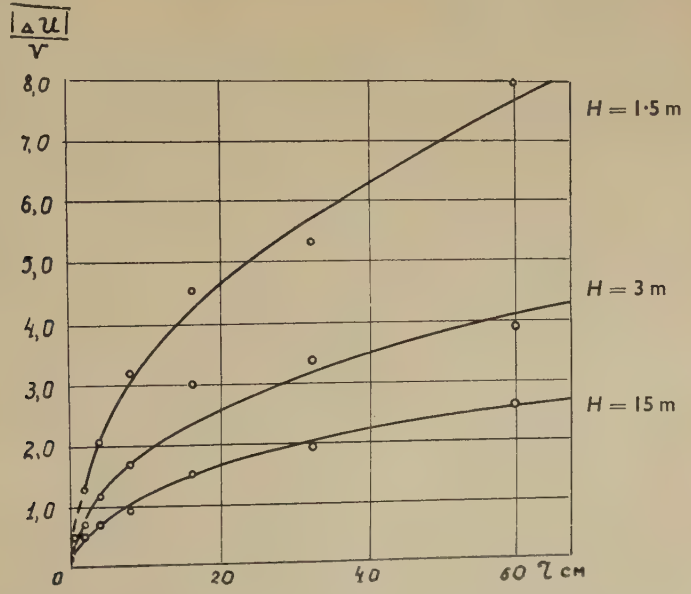


Figure 3. The empirical structure functions of the field of wind velocity.

The data of the measurements from the hot-wire anemometer and microthermometer were recorded with mirror galvanometers having a time lag of the order of 0·005 sec (Fig. 2 shows the patterns of the records obtained). These records were then used to obtain the mean absolute values. The transition from the mean absolute value to the r.m.s. is realized through multiplication by a number, determined from empirical distribution functions.

If we measured the difference between the wind velocity or temperature at two points, we obtained the value of the corresponding structure function $D(r)$ for some fixed r. Fig. 3 (taken from Obukhov 1951) shows examples so obtained of the lateral structure functions of the wind velocity at different heights above the surface (more exactly, examples of the functions $d(r) = \overline{|u_n(M) - u_n(M')|}$, which is here proportional to $\sqrt{D_{nn}(r)}$). The continuous lines here correspond to the theoretical '$\frac{2}{3}$ law.' The results relating to different heights z and different values of the mean wind velocity V (i.e. to different values of the friction velocity V_*) can be described here with the following theoretical formula :

$$D_{nn}(r) = C_2 \frac{V_*^2}{(kz)^{\frac{2}{3}}} r^{\frac{2}{3}},$$

where C_2 is a numerical constant of the order of a unity. The empirical data lead in this case to the values $C_2 = 1{\cdot}2$. An example of the structure function of the temperature field obtained by Krechmer (1952) is given in Fig. 4. Later on the methods for the determination of the structure function of the temperature field have been improved by Tatarskiy (1956a). He replaced photographic records for the determination of the absolute or r.m.s. values by a special radio scheme, enabling him to record the mean square of the current in a diagonal of the bridge. He was thus able to obtain in a short period of time about 400 measurements of the structure function of the temperature field in different places and at different heights above the surface and under different meteorological conditions. The results of the measurements completely proved the theoretical '$\frac{2}{3}$ law' and cleared up the question of the dependence of the coefficient A_4 in formula (9) upon the meteorological conditions (see Fig. 5 in which the empirical values of the coefficient $B = \sqrt{A_4}$ are presented in the form of the function of $z^{-\frac{1}{3}} T_*$, where z is the height of the observation point above the earth, and $T_* = \dfrac{T(z_1) - T(z_2)}{\log z_2 - \log z_1}$ is the coefficient characterizing the profile of the mean temperature $T(z)$ in the surface layer).

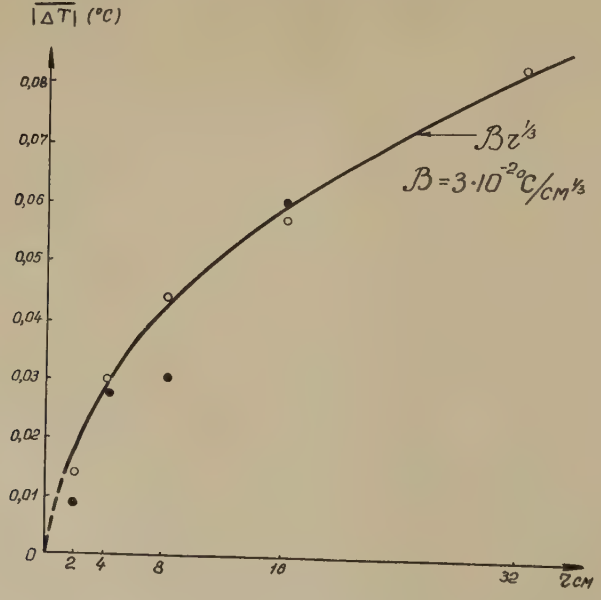


Figure 4. The empirical structure function of the temperature field.

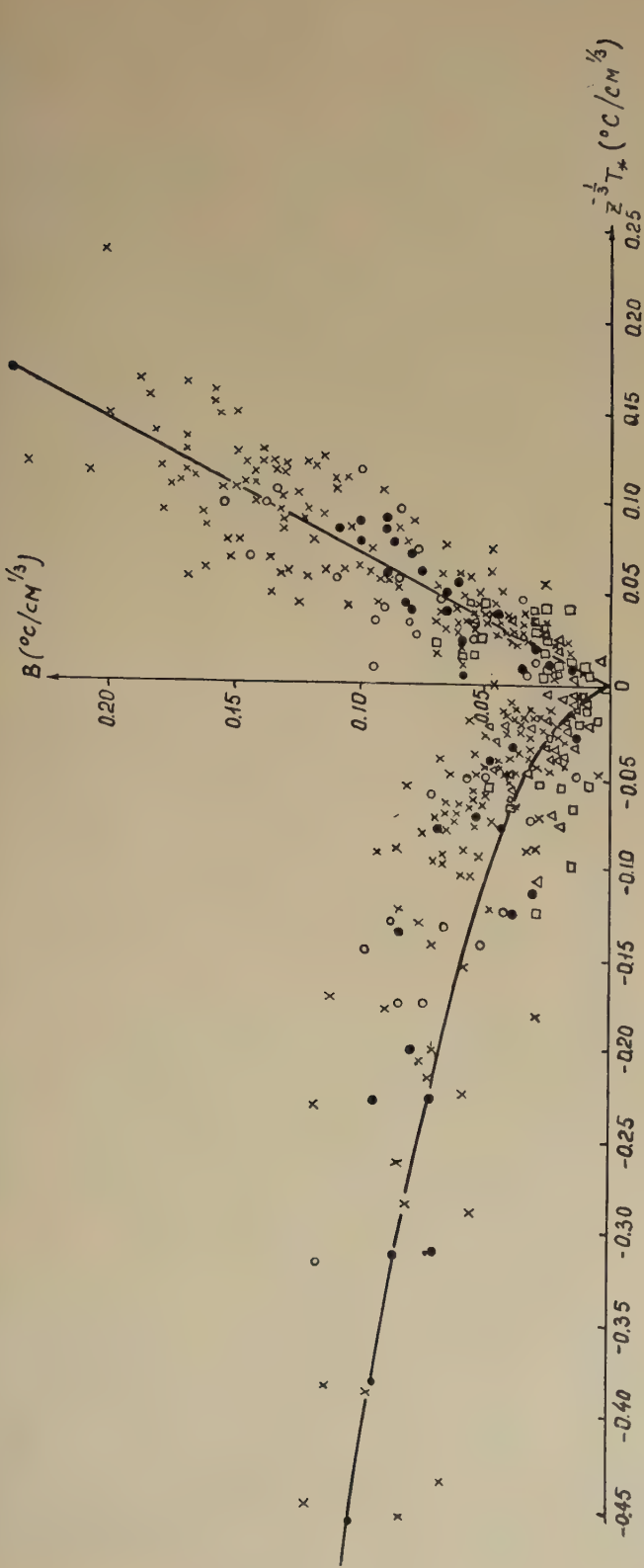


Figure 5. The dependence of the structure characteristic $B = \sqrt{A_4}$ of the temperature field on the meteorological conditions.

- ● August 1954.
- △ March-April 1955.
- ○ June-July 1955, $z = 1{\cdot}5$ m.
- □ June-July 1955, $z = 22$ m.
- × July-September 1955.

5. Application of results

The results given above on the microstructure of the meteorological fields have various applications. For example, in the study of the influence of turbulent fluctuations on the readings of meteorological instruments, we have to solve differential equations on the right side of which there stands the function $u(t)$ representing the measured quantity. Because of turbulence the quantity has random fluctuations so that the function $u(t)$ has a shape similar to that shown in Fig. 2. Naturally, we cannot expect a solution of this equation in an analytical form but must confine ourselves to the determination of the simplest statistical characteristics of the solution (first of all its mean value and variance) if the statistical characteristics of the function $u(t)$ are given. Such an analysis for some of the most important instruments (for rotation anemometers and thermometers) was made by Yaglom (1954). This analysis was based on the consideration of the time structure function

$$D(\tau) = \overline{[u(t) - u(t-\tau)]^2}, \qquad . \quad . \quad . \quad (11)$$

which can be expressed approximately through the space structure function (2) supposing that the disturbances in the atmosphere move with the mean wind. This investigation gave, in particular, the excess of an anemometer reading in the real atmosphere over the actual wind velocity*. Particularly, the difference δ between the mean value of the anemometer reading and the actual mean wind V is given by

$$\delta = \frac{1+\alpha}{L} \int_0^\infty e^{-\frac{\tau V}{L}} D(\tau)\, d\tau \qquad . \quad . \quad . \quad (12)$$

where L is the so-called 'synchronization length' of the rotation anemometer which characterizes its time lag and is of the order of 10-20 m in the case of the ordinary cup anemometer and α is a non-dimensional constant which is different for different anemometers but usually is noticeably less than unity. In particular, when

$$D(\tau) = C\tau^\beta, \quad 0 < \beta < 2 \qquad . \quad . \quad . \quad (13)$$

it follows from (11) that

$$\delta = \frac{(1+\alpha)\,\Gamma(1+\beta)}{2} C \frac{L^\beta}{V^{\beta+1}} = \frac{(1+\alpha)\,\Gamma(1+\beta)}{2} \frac{D(L/V)}{V}, \qquad . \quad (14)$$

where Γ denotes the Γ-function.

Another problem, more important than the above, involving data on the microstructure of meteorological fields, is connected with the study of wave propagation in the earth's atmosphere as affected by small fluctuations of the refractive index. For light waves the inhomogeneities of the refractive index are mainly due to turbulent fluctuations of the temperature; for sound, additionally, one has to take into account the influence of the fluctuations of the wind, and in radiophysical problems the humidity fluctuations play an essential role. It is important therefore to relate the microstructure of the field of refractive index with the microstructure of the main meteorological fields.

The principal results of the theory of wave propagation in a medium with random inhomogeneities of the refractive index, obtained in recent years by the Soviet authors are set forth in a review by Krasilnikov and Obukhov (1956). These results express the statistical characteristics (the mean square values, correlation and structure functions) of fluctuations of amplitude, A, and phase, S, of a wave of wavelength, λ, propagated in a turbulent layer of given thickness, L, in terms of the correlation or structure function of the

* The fact that the rotation anemometer because of its inertia always shows too high a wind is well known (for example, see Middleton and Spilhaus 1953, p. 146-147). A number of theoretical and experimental works were devoted to the study of this phenomenon. However, in all these works one treated an artificially gusty wind with wind velocity changing periodically either with discrete jumps or according to a sinusoid. It is clear that both these models present a very rough approximation to conditions in the real atmosphere.

refractive index. For example, if the refractive index $n(M)$ is $n(M) = 1 + \mu(M)$, where μ characterizes the small random inhomogeneties and the correlation function of $\mu(M)$ is equal to

$$B(r) = \overline{\mu(M)\,\mu(M')} = B\,e^{-r^2/2l} \quad . \quad . \quad . \quad (15)$$

then for the mean square of $\log \frac{A}{A_0}$ where A_0 is a mean value of A, the following formula has been obtained

$$\overline{\left(\log \frac{A}{A_0}\right)^2} = (2\pi)^2 B \frac{lL}{\lambda^2}\left(1 - \frac{1}{D}\operatorname{arc\,tg} D\right), \quad D = \frac{\lambda L}{2l} \quad . \quad . \quad (16)$$

The formula for the mean square of the phase fluctuation differs from that of (15) only in sign : minus in the right side is changed into plus (see Obukhov 1953). The result (15) is noticeably simplified in the limiting case when $D \ll 1$ (' the geometrical optics approximation ') or when $D \gg 1$. Another example has been considered by Tatarskiy (1956b) who has assumed that the refractive index $n(M)$ is dependent only on the temperature and therefore satisfies the ' $\frac{2}{3}$ law ' :

$$\overline{[n(M) - n(M')]^2} = D_{nn}(r) = A_5\, r^{\frac{2}{3}} \quad . \quad . \quad . \quad (17)$$

By this assumption he has obtained the relation

$$\overline{\left(\log \frac{A}{A_0}\right)^2} = 10{\cdot}2\, A_5\, \lambda^{-\frac{7}{6}} L^{\frac{11}{6}} \quad . \quad . \quad . \quad . \quad (18)$$

Using these results we obtain estimates of the values characterizing the scintillation of the stars (scintillation amplitude, intensity of ' star dancing,' sizes of stellar-shadow bands). These estimates are in agreement with the published observations (it is regrettable they are still very few) if the effective thickness of the layer of the atmospheric disturbances is of the order of 2-3 km and the value of the structure characteristic $B = \sqrt{A_4}$ for the temperature field is of about 8×10^{-3} °C/cm$^{\frac{1}{3}}$.

The observations of stellar scintillation, properly organized, would provide valuable information about the statistical characteristics of the micro-inhomogeneities of the atmosphere.

The radiophysicists are interested first of all in the evaluation of the intensity of the scattered field arising from the micro-inhomogeneities of the atmosphere. Many investigations have recently been devoted to this problem. We shall refer here only to the recent work of Silverman (1956) which states that the data on the microstructure of meteorological fields (the ' $\frac{2}{3}$ law ' and the empirical data obtained by Krechmer shown in Fig. 4) provide reasonable estimates of the intensity of the scattered field which agree with direct measurements of radio signals beyond the horizon.

References

Corrsin, S. 1951 ' The decay of isotropic temperature fluctuations in an isotropic turbulence,' *J. App. Phys.*, **22**, No. 4, pp. 469-473.

Gödecke, K. 1936 ' Messungen der Atmosphärischen Turbulenz,' *Ann. Hydrogr.*, **64**, Pt. 10, p. 469.

Heisenberg, W. 1948 ' Zur statistischen Theorie der Turbulenz,' *Zeitr. Phys.*, **124**, No. 7, pp. 628-657.

Inoue, E. 1950 ' On the temperature fluctuations in a heated turbulent fluid,' *Geophys. Notes*, Tokyo Univ., **3**, No. 34.

Keller, L. and Friedman, A. 1924 ' Differentialgleichungen für die turbulente Bewegung einer kompressibken Flüssigkeit,' *Proc. First Inter. Congr. App. Mech.*, Delft, pp. 395-405.

Kolmogorov, A. N. 1941a 'The local structure of a turbulence in incompressible fluids at large Reynolds numbers,' *Dokl. Akad. Nauk S.S.S.R.*, **30**, No. 4, pp. 299-303.

1941b 'The scattering of energy in locally isotropic turbulence,' *Ibid.*, **32**, No. 1, pp. 19-21.

Krasilnikov, V. A. and Obukhov, A. M. 1956 'On the propagation of waves in a medium with random inhomogeneities of the coefficient of refraction,' *Akust. zhum.*, **2**, No. 2, pp. 107-112.

Krechmer, S. I. 1952 'An investigation of the micropulsation of the temperature field in the atmosphere,' *Dokl. Akad. Nauk. S.S.S.R.*, **32**, No. 1, pp. 55-58.

1954 'The methodology of measuring the micropulsation of wind speed and temperature in the atmosphere.' *Akad. Nauk. S.S.S.R., Trud. Geofiz. Inst.*, No. 24 (151), pp. 43-111.

MacCready, P. B. 1953 'Atmospheric turbulence measurements and analysis,' *J. Met.*, **10**, No. 5, pp. 325-337.

Middleton, W. B. K. and Spilhaus, A. F. 1953 *Meteorological instruments*, Univ. of Toronto Press, pp. 146-147.

Obukhov, A. M. 1941 'On the distribution of energy in the spectrum of turbulent flow,' *Izv. Akad. Nauk. S.S.S.R.* Geograph. and Geophys. Series No. 4, pp. 453-466.

1942 'On atmospheric turbulence,' *Ibid.*, Phys. Series, **6**, No. 1, pp. 59-63.

1949a 'The structure of the temperature field in turbulent flow,' *Ibid.*, Geograph. and Geophys. Series, **13**, No. 1, pp. 58-69.

1949b 'The pulsation of pressure in turbulent flow,' *Dokl. Akad. Nauk. S.S.S.R.*, **66**, No. 1, pp. 17-20.

1951 'Characteristics of the microstructure of the wind in the surface layer of the atmosphere,' *Ivz. Akad. Nauk. S.S.S.R.*, Geophys. Series No. 3, pp. 49-68.

1953 'On the influence of weak inhomogeneities in the atmosphere on the propagation of sound and radiation,' *Ibid.*, Geophys. Series No. 2, pp. 155-165.

Obukhov, A. M. and Yaglom, A. M. 1951 'The microstructure of turbulent flow,' *Prikl. matem. i. mekh.*, **15**, No. 1, pp. 3-26.

Shiotani, M. 1950 'Turbulence in the lowest layers of the atmosphere,' *Sci. Rep.*, Tokyo Univ., Series 5, Geophys., **2**, No. 3.

1955 'On the fluctuations of the temperature and turbulent structure near the ground,' *J. Met. Soc.*, Japan, **33**, No. 3, pp. 117-123.

Silverman, R. A. 1956 'Turbulent mixing theory applied to radio scattering,' *J. App. Phys.*, **27**, No. 7.

Tatarskiy, V. I. 1956a 'The microstructure of the temperature field in the surface layer of the atmosphere,' *Izv. Akad. Nauk. S.S.S.R.*, Geophys. Series, No. 6, pp. 689-699.

1956b 'On the pulsations in the amplitude and phase of waves which are being propagated in a weakly inhomogeneous atmosphere,' *Dokl. Akad. Nauk. S.S.S.R.*, **107**, No. 2, pp. 245-248.

Yaglom, A. M. 1949a 'On the local structure of the temperature field in turbulent flow,' *Ibid.*, **69**, No. 6, pp. 743-746.

1949b 'On the field of acceleration in turbulent flow,' *Ibid.*, **67**, No. 5, pp. 595-798.

1954 'On allowing for the inertia of meteorological apparatus in measurements in a turbulent atmosphere,' *Akad. Nauk. S.S.S.R., Trud. Geofiz. Inst.*, No. 24 (151), pp. 112-162.

551.594.18 : 551.551.21

Atmospheric electrical agitation

By H. ISRAËL

*Meteorologisches Observatorium des Deutschen Wetterdienstes, Aachen**

(Manuscript received 5 May 1958; in revised form 1 Nov. 1958)

SUMMARY

The paper shows the so-called ' agitation ' of the atmospheric electrical elements in relation to the atmospheric turbulence, and demonstrates in a series of examples the behaviour of the agitation and its application to the analysis of atmospheric electrical processes. We can consider it as a further element in the coupling between atmospheric electricity and meteorology. It appears that the central significance of the coupling can be seen in the various atmospheric exchange processes and their different electrical effects.

1. THE AGITATION AS A GEOPHYSICAL PHENOMENON

All data of a geophysical nature, because they include permanent fluctuations, need to be averaged over a definite period of time. This holds for the averaging over a definite long period of time – year, month, day – as much as for the averaging over the period of duration of the measuring process itself, since each individual measurement, too, proves when analysed to be composed of different values varying around an average†.

We denote these properties of a geophysical element as its variability or agitation and we understand by ' variability ' its slow fluctuation and by ' agitation ' its fast fluctuation, and we will attempt to understand them. As typical examples of an agitation we know the atmospheric pressure oscillation, the wind gustiness, the temperature variation; and from their behaviour during the course of a day or in relation to a definite weather cycle we derive an expression for the atmospheric mass-exchange.

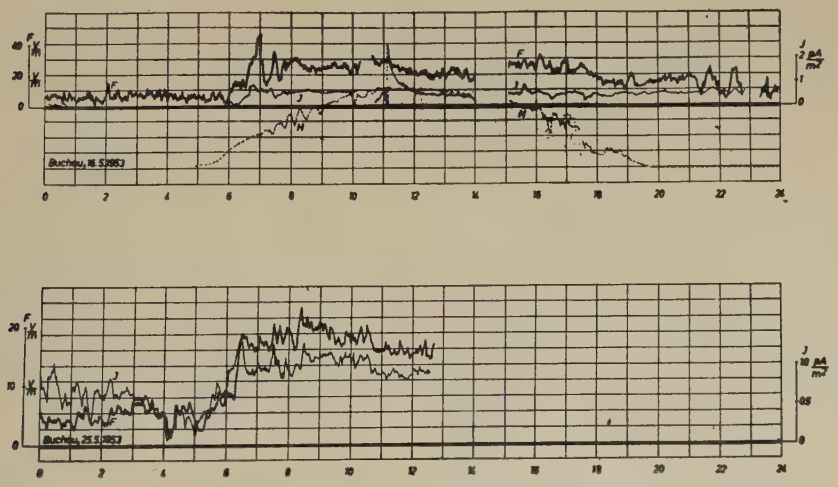


Figure 1. Record of the potential gradient (heavy line curve F); air-earth current density (faint line curve I) and brightness (broken line curve H) at Buchau a.F. on 16 and 25 May 1953 (Israël 1958).

† We mean of course, the deviations arising for purely geophysical reasons and not those arising from experimental errors.

* The investigations reported in this paper are based partially on researches sponsored by the generous help of the Geophysics Research Directorate of the U.S. Air Force Cambridge Research Center, Air Research and Development Command, United States Air Force, under Contract No. AF 61 (514)-640.

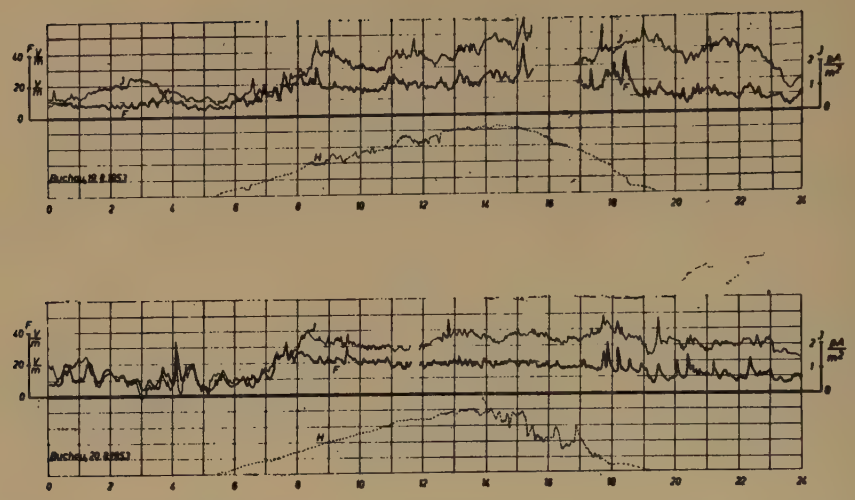


Figure 2. Record of the potential gradient (heavy line curve F), air-earth current density (faint line curve I) and brightness (broken line curve H) at Buchau a.F. on 19 and 20 Aug. 1953.

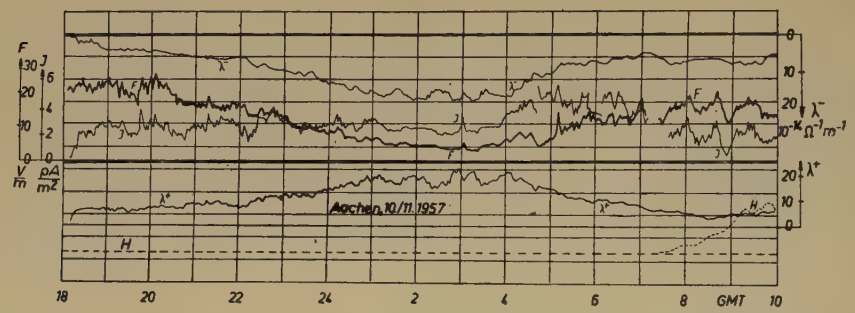


Figure 3. Records of the potential gradient (heavy line curve F), air-earth current density (faint line curve I) the positive polar and negative polar conductivity (λ^+ and λ^-) and the brightness (broken line curve H).

In the field of atmospheric electricity we find similar phenomena; for example, records of the atmospheric electrical potential gradient obtained by Nordman (1911) and by Kasemir (see Israël 1957b) show in all cases a strong agitation superimposed on the course of the diurnal variation, increasing with sunrise, reaching a maximum about mid-day, and falling off towards the evening.

Fig. 1 shows records of the potential gradient, air-earth current density and local brightness (as displayed by a horizontally-lying photocell). Besides the diurnal variation of the agitation one notices also the almost complete parallelism between the potential gradient and air-earth current density.

Fig. 2 shows two records of a similar nature on 19 and 20 Aug. 1953, in which a few other influences can be recognized : whereas on 19 Aug. the agitation clearly increases from the night towards the day, it remains the same also during the following night.

Fig. 3 shows a record at Aachen, of potential gradient, air-earth current density and both the polar conductivities. One notices a relationship between the conductivities on the one hand and the agitation of the field and of the current on the other, in such a way that with increasing conductivity the agitation appears to decrease and vice versa.

2. The agitation as a characteristic factor in atmospheric electricity

(a) *Definition*

The examples so far mentioned only partially demonstrate the whole complexities of the atmospheric electrical fluctuations. First, they are all taken without exception at times when there was no precipitation, fog, or other influences which affect atmospheric electricity such as thunderstorms or smoke disturbances. This selection is due to the aim which we set at the beginning : all occasions with hydrometeors present show atmospheric electrical effects so clearly that the classification of cause and effect is evident and any attempt to separate local and global influences is merely superfluous. Therefore we have to investigate the influences arising from hydrometeors separately and will not consider them here. (see e.g., Simpson 1949; Chalmers 1957). Secondly, we want to limit ourselves by defining the word ' agitation ' as covering only a definite periodic-range of the atmospheric electrical fluctuations, whereas in reality oscillations of *all* periods are to be expected and do exist. We will separate the fluctuations of the total fluctuation spectrum which occur within a duration of more than 1 to 2 minutes to one hour, and study them in more detail. The limitation of the lower time limit to 1 or 2 minutes is naturally arbitrary, but it is certainly necessary to eliminate the differences in the lag between each individual measuring instruments, which would otherwise lead to different data on the agitation. The limitation to the greater time limit is because variations in the different hourly-mean values are automatically accounted for in the usual evaluation, while variations within this ' fundamental unit ' of one hour are not.

Special attention, therefore, should be paid to this region of atmospheric electricity which has been so far very neglected and more so, because in this region meteorological factors have a very special behaviour.

The limited region mentioned becomes all the more important because it covers that range of time in which the transition takes place from stationary to non-stationary processes in atmospheric electricity.

(b) *Frequency and amplitude of the agitation*

Statistics of predictions about the agitation must show the magnitude of the individual fluctuations, their duration and their number. For various reasons there are certain difficulties in obtaining these quantities. To begin with, one must naturally make the data independent of the lag of each measuring instrument and its method of operation, i.e., only those variations which can be recorded without distortion by the measuring equipment are to be valued as agitation. As a first step towards this, the range of agitation has been limited to periods of more than 1-2 minutes.

One can gather the data about the agitation either from records such as Figs. 1-3, or ascertain them from the variograph, which, as the name indicates, shows only the variations with time and not the absolute value. Both methods have their own advantages and drawbacks and it is difficult to prefer one to the other. For example, when the agitation is taken from the normal records, one gets in one evaluation process alone the hourly average value of the factor concerned, its absolute and its relative scattering-band width. The variograph shows up the agitation more magnified and in finer details, but not necessarily without distortion (Israël 1939).

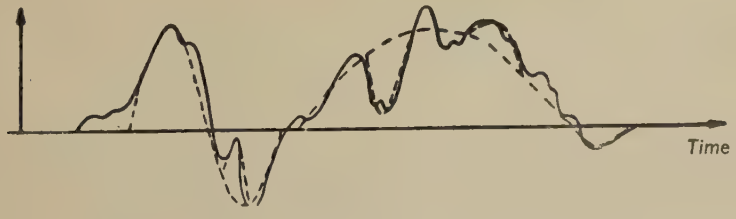


Figure 4. Definition of the agitation (Courvoisier 1951).

As suggested by Courvoisier (1951) one can look in the recorded curves for parts which are sine-wave or approximately so (as shown by the broken line in Fig. 4) and classify these according to their period and amplitude. Such a method gives a comparatively reliable picture of the agitation. However, for practical application where one needs a continuous evaluation, this method would need a lot of time.

To make matters easy the following method is used : taking an hour as the required duration of time, the number of times the curve reverses in this interval, gives us a relative measure of the average length of the period. These data which are sufficiently easy and quick to arrive at, we will denote as ' frequency of the agitation.' For the amplitude of the fluctuations we make use of a definition given by Bidlingmeyer (1913) for the magnetic agitation, which has been found to be suitable for the atmospheric electrical agitation also (Kuhn 1933), and denote the difference between the extreme values within one hour as the ' amplitude of the agitation.'

Another recent definition put forward by Whitlock and Chalmers (1956) is based on the minute as the time unit and denotes the ' field unrest ' as the average change from one minute to the next, expressed as a percentage of the minute average concerned. To derive this, field-mill recordings are taken.

3. Results : The phenomenology of the agitation

(*a*) *Statistics*

(i) Material

The results discussed below arise mainly from the experiments on atmospheric electricity carried out by us at different places on the Swiss Alps during the year 1954-55 (Israël 1957c). The recordings obtained there are evaluated with respect to the agitation only for a part of a year; however, the values obtained may suffice to show a first glimpse of the phenomenology of the agitation.

Conforming to the definitions already given, for each hourly range, the average values of the potential gradient and air-earth current density, and the difference between the highest and lowest values and the reversal points of the recordings, were determined. The investigation is confined to the potential gradient and air-earth current density since, because of experimental difficulties, it is impossible to deal with conductivity in this sense ..

(ii) Average values

Table 1 gives an initial survey. The data, average and extreme values, are valid for the period from November 1954 to May 1955 for Jungfraujoch; March to May 1955 for Gornergrat, and from December 1954 to June 1955 for Payerne.

The Table shows that the average hourly values of the agitation of potential gradient and air-earth current density have amplitudes which are of same order of magnitude as the mean hourly values of the elements themselves. The lower part of the Table shows records that change their tendency on the average about 2·5 to 9 times per hour, or once in about 6·5 to 24 minutes.

The table shows, furthermore, that the ' relative amplitude of agitation ' (i.e., the amplitude of fluctuation covered during one hour divided by the hourly mean of the element itself) is *markedly* greater both in average and extreme values at the flat-land station of Payerne than at the high-mountain stations of Jungfraujoch and Gornergrat. In other words, the general character of the temporal variation of field and current is more quiet at elevated stations than in flat-land situations; more so for current than for potential gradient. The frequency of agitations also displays a similar behaviour; at high-altitude stations the frequency is less than at low-lying stations; this behaviour is more pronounced, also, for air-earth current density than for the field.

TABLE 1. AVERAGE AND EXTREME VALUES OF THE AGITATION OF POTENTIAL GRADIENT AND AIR-EARTH CURRENT DENSITY AT JUNGFRAUJOCH (JJ), GORNERGRAT (GG) AND PAYERNE (Py) (MONTHLY AVERAGES). THE AMPLITUDES OF THE AGITATION ARE GIVEN IN EACH CASE AS THE PERCENTAGE OF THE AVERAGE VALUE FOR POTENTIAL GRADIENT AND AIR-EARTH CURRENT DENSITY RESPECTIVELY.

	Potential gradient (percentage)				Air-earth current density (percentage)		
Station	JJ*	GG†	Py‡	Station	JJ*	GG†	Py‡
Amplitude				Amplitude			
Mean	39	37	68	Mean	35	23	86
Maximum	81	39	86	Maximum	69	25	114
Minimum	21	33	50	Minimum	22	21	57
Frequency				Frequency			
Mean	4·9	7·8	7·6	Mean	2·5	2·8	9·2
Maximum	9·0	10·8	10·6	Maximum	5·1	3·5	11·6
Minimum	2·3	6·0	4·5	Minimum	1·7	2·0	5·8

* Period from Nov. 1954 to May 1955. † Period from Mar. 1955 to May 1955. ‡ Period from Dec. 1954 to June 1955.

When one compares the values given here with those obtained by the author (1943) from field-variograph records, and with the values given by Whitlock and Chalmers (1956) for the agitation, one notices very strikingly the larger field-agitation in the case considered now; the ' variographic agitation ' is about one or two per cent of the value of the potential gradient. For *fine* weather with cloudless skies, Whitlock and Chalmers find an average value of 2·9 per cent; values which come close to those given in Table 1. The reason why these values do not exactly agree lies in the fact that different methods are used for recording or calculation. Because of its period sensitivity the variograph strongly accentuates the shorter ' minute fluctuations ' and suppresses the longer ones : while the figures in Table 1 stress the longer fluctuations and suppress the shorter. Whitlock and Chalmers procedure puts the emphasis in between the values of the variograph and the method which we have used here.

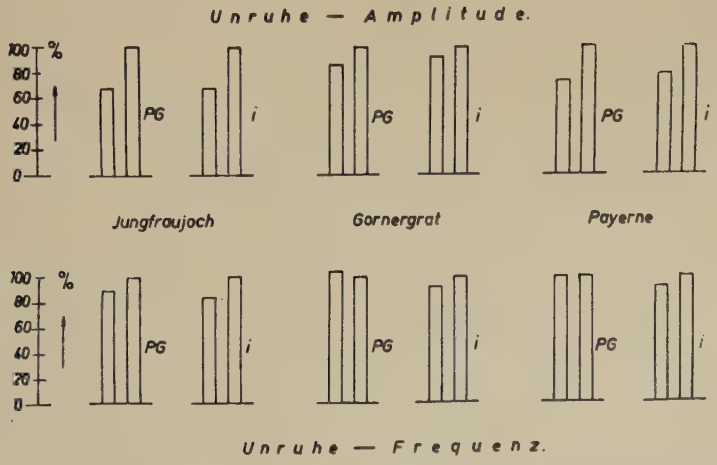


Figure 5. Comparison of the average values of the ' agitation amplitude ' (upper curve) and ' agitation frequency ' (lower curve) of the potential gradient and air-earth current density at Jungfraujoch, Gornergrat and Payerne. The right column gives in each case the average value of the total material (100 per cent set) and the left column the average value of the ' undisturbed days ' as a fraction of the total average.

(iii) Agitation and the weather conditions

According to its definition, the agitation comes in the range of non-stationary processes s and therefore belongs to the phenomena of local atmospheric electricity and, as is shown by experience, it is closely related to the weather.

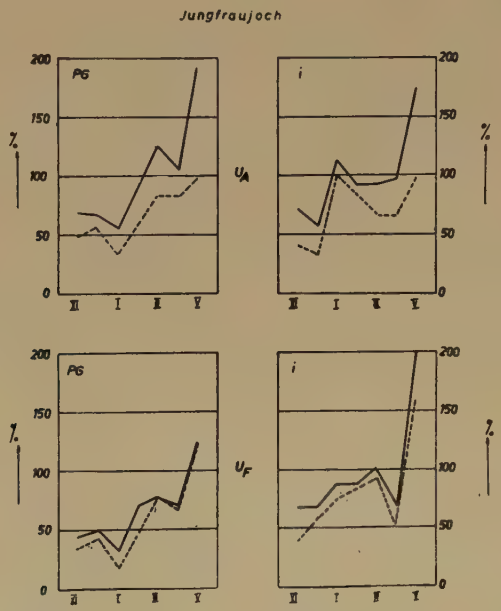


Figure 6. Annual variation of the agitation of potential gradient and air-earth current density at Jungfraujoch for the period from November to May. Upper part of the figure: agitation-amplitude. Lower part: agitation-frequency. The monthly average of the total material (full line curve) and that of the undisturbed days (broken line) are given as percentages of the total average.

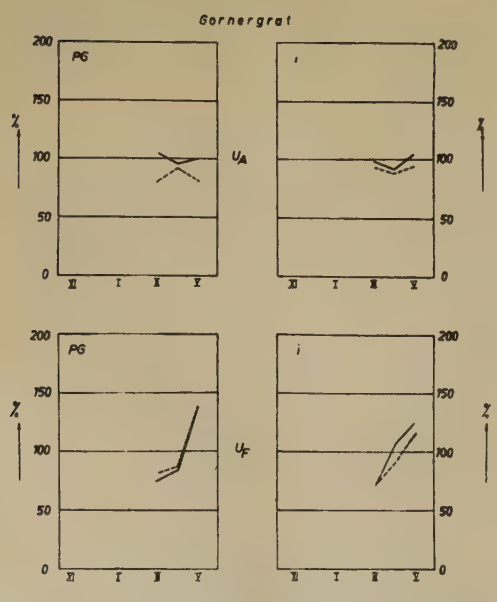


Figure 7. The same as Figure 6 for Gornergrat during the period from March till May.

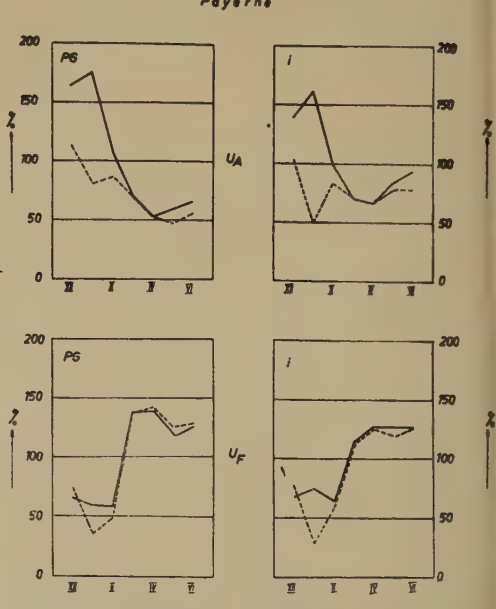


Figure 8. The same as Figure 6 for Payerne during the period from December till June.

If, out of the complete material one picks out the so-called ' undisturbed days ' and compares the average values of the agitation in these two groups, one gets the result shown in Fig. 5.

One notices that on *undisturbed days* the average values of the agitation amplitude and agitation frequency are less, sometimes considerably less (more than 30 per cent), than the corresponding values of the *whole material*. (In the case of agitation frequency there is one exception). By ' undisturbed days ' are meant days with little or no cloud formation and only slight air-motion, as are often met with in high-pressure areas. On such days the agitation does not disappear, rather it approaches a certain minimum figure which can be up to 30 per cent below the average values for the times which are completely free of fog and precipitation. With the worsening of the weather, increased cloud formation, wind, etc., the agitation parameters systematically increase. Hence it is statistically proved that the weather affects the agitation. In order to give a closer analysis, an attempt was made to find the statistical influence of individual weather factors. The results proved to be unsatisfactory, showing that the material available is still too little or that the investigation method is not yet adapted for such a detailed investigation. For individual examples about the influence of the weather, see below.

(iv) Annual variation of the agitation

As shown in Figs. 6 to 8, the agitation of the potential gradient and air-earth current density undergo systematic variations during the course of a year.

These figures show that the result of Fig. 5 remains valid for the individual monthly values also. Further, one notices that in all the cases the agitation changes considerably from winter to summer. At both the high-altitude stations the amplitude and frequency increase about three times. At Payerne however, the amplitude and frequency behave exactly dissimilarly to each other; while the frequency increases as at the high-altitude stations, the amplitude falls to about one-third of its value in winter.

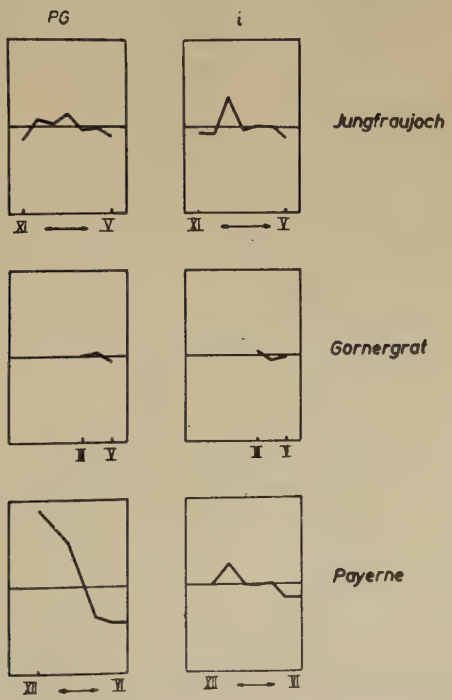


Figure 9. The average values of potential gradient and air-earth current density for the periods of November to May at Jungfraujoch, March to May at Gornergrat and December to June at Payerne.

A further peculiarity of the annual variations is that they do not change uniformly, rather – especially in the case of frequency – there seem to be more or less distinct jumps from the values at winter to the summer values. At the high-altitude stations these 'jumps' occur in April to May, this is about one or two months later than at Payerne, where the fluctuations are already pronounced in February to March.

The last symptom, especially the opposite behaviour of the agitation amplitude and frequency at Payerne from winter to summer, points to the fact that the agitation parameter evidently depends on several different factors which are different in their effects. For example, if one takes for comparison the average values of the elements themselves, i.e., potential gradient and air-earth current density (Fig. 9) the following points come forth.

At the high-altitude stations, where the average values of potential gradient and air-earth current density do not change remarkably, the amplitude and frequency of the agitation increases from winter to summer. On the other hand at Payerne, where the potential gradient and air-earth current density decrease the agitation amplitude also decreases, but at the same time its frequency increases.

This leads to the conclusion that evidently both the agitation parameters undergo a clear annual cycle, with higher values in summer and smaller values in winter. Moreover the agitation amplitude seems to be further related to the value of the element itself which is superimposed on its annual cycle.

Additionally, Table 2 shows side-by-side the average values of the absolute and relative agitation amplitudes for the total material and for that of the undisturbed days*. From this one notices that the relative agitation amplitude behaves, as a rule, similarly to the absolute values. Only in the case of the potential gradient at Payerne does there appear to be an anomaly, in such a way that the absolute agitation amplitude decreases from winter to summer whereas the relative agitation increases.

TABLE 2. MONTHLY AVERAGE VALUES OF THE ABSOLUTE AND RELATIVE AGITATION AMPLITUDE : U_{abs} and U_{rel} OF THE POTENTIAL GRADIENT AND AIR-EARTH CURRENT DENSITY AT THE STATIONS OF JUNGFRAUJOCH, GORNERGRAT AND PAYERNE

		Potential gradient		Air-earth current density	
		U_{abs} (volt)	U_{rel} (per cent)	U_{abs} (10^{-12} A)	U_{rel} (per cent)
		Jungfraujoch			
1954	Nov.	138 (99)	28 (19)	20 (12)	26 (14)
	Dec.	138 (114)	24 (23)	16 (9)	22 (15)
1955	Jan.	113 (96)	21 (14)	32 (28)	29 (26)
	Feb.	179	30	26	32
	Mar.	250 (170)	49 (30)	26 (19)	31 (21)
	Apr.	211 (167)	41 (30)	27 (19)	33 (20)
	May	382 (194)	81 (43)	49 (28)	69 (37)
		Gornergrat			
1955	Mar.	95 (75)	38 (30)	2·2 (2·1)	21 (19)
	Apr.	88 (85)	33 (31)	2·0 (2·2)	22 (20)
	May	92 (75)	39 (31)	2·3 (2·5)	25 (22)
		Payerne			
1954	Dec.	101 (70)	66 (54)	2·5 (1·9)	114 (106)
1955	Jan.	107 (49)	78 (34)	2·9 (0·9)	112 (129)
	Feb.	65 (53)	52 (51)	1·8 (1·5)	83 (68)
	Mar.	44 (44)	50 (47)	1·3 (1·3)	65 (62)
	Apr.	33 (33)	64 (59)	1·2 (1·2)	57 (57)
	May	37 (29)	76 (59)	1·5 (1·4)	79 (88)
	Jun.	40 (34)	86 (67)	1·7 (1·4)	94 (70)

(In each group, the left column gives the values averaged from the total material, and the right column the values which are valid for the undisturbed days).

* 'Relative' agitation amplitude of an hour is defined as the quotient of the amplitude divided by the hourly mean of the element itself.

(v) The diurnal variation of the agitation

Considering the diurnal variation of both the agitation parameters, some further characteristics can be seen, as shown in Figs. 10 to 12.

Because of lack of sufficient data, it is not yet possible to subdivide here into periods of a month or a season, hence the figures give only the mean diurnal variation for the total period, and that too, once for the total material and once for the undisturbed days.

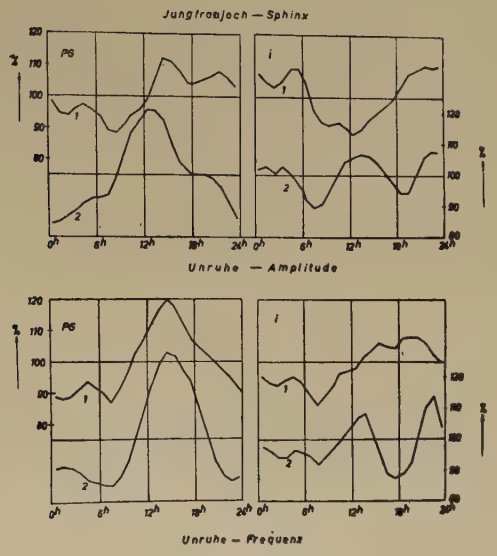


Fig. 10. Average daily cycle of the agitation amplitude (top) and the agitation frequency (bottom) of potential gradient (left) and air-earth current density (right) for Jungfraujoch-Sphinx, November till May (percentage representation). Curve 1. Total material. Curve 2. Mean of the undisturbed days.

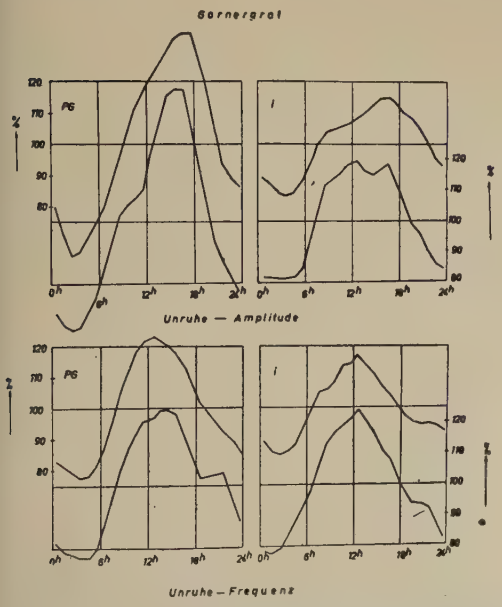


Figure 11. The same as Figure 10 for Gornergrat, March till May.

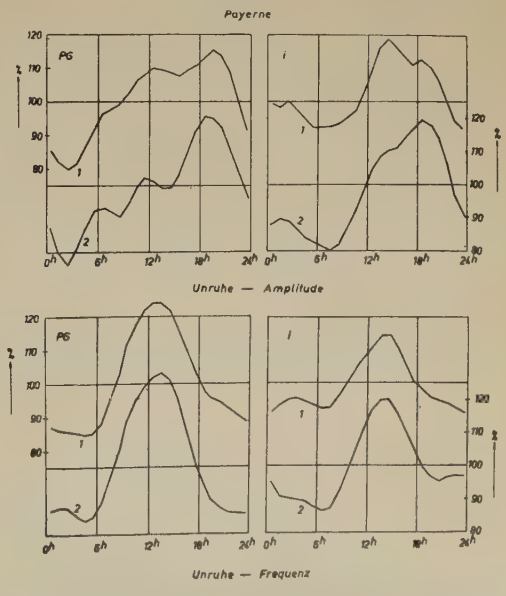


Figure 12. The same as Figure 10 for Payerne, December till June.

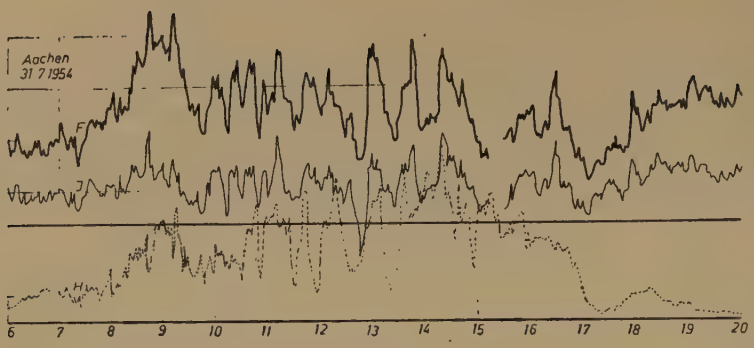


Figure 13. Recordings of the potential gradient (F), air-earth current density (I) and brightness (H) at Aachen on 31 July 1954.

To get more information about typical behaviour from very strongly fluctuating diurnal variations the curves are smoothed according to the equation :

$$\bar{x}_n = \frac{x_{n-2} + 2x_{n-1} + 3x_n + 2x_{n+1} + x_{n+2}}{9}$$

Figs. 10 to 12 show that for potential gradient the agitation parameters have a closely similar diurnal variation. The agitation amplitude and frequency jointly display a tendency to increase in value from night to daytime, and from the evening onwards this seems to retrace. Apart from minor differences this seems to hold good both for the total material and for the undisturbed days. At Gornergrat and Payerne the air-earth current density displays a similar behaviour; only at Jungfraujoch does this relation seem to be less clear.

(*b*) *Individual cases*

Considering the different atmospheric electrical factors individually, it will be seen that every meteorological process shows itself in the agitation. For example, it became clear a long time ago that frontal zones and inversions – so far as they occur without precipitation or fog (only such cases have to be considered here) – are distinguished by particularly strong agitation of the atmospheric electrical elements (Israël 1943; Rossmann 1950; Koenigsfeld 1955).

The behaviour of the atmospheric electrical factors during times of cloudiness offers another example of their agitation. If potential gradient, air-earth current density, and, additionally, the general local brightness are recorded one will see (as was first indicated by Dolezalek and Fries 1956) that the changes in potential gradient and air-earth current density run closely parallel to changing brightness below the clouds. (Fig. 13 shows an example).

For individual fine-weather cumulus cloud, Lecolazet (1946) has shown the direct effect of the charges contained in the clouds on the agitation of potential gradient.

4. On the interpretation

(*a*) *Agitation and mean level of the element concerned*

The experiences pictured so far help to clarify the basic decisive questions, whether the agitation is characteristically connected with the value of potential gradient and air-earth current density, or an additional property and factor which is independent of the mean value. Both these cases seem to be possible a priori.

When we plot the agitation amplitude in relation to the value of the element concerned the results appear as shown in Fig. 14. The figure shows that there is evidently a definite relation between the agitation-amplitude and the element concerned, but this fluctuates very strongly, and it differs from place to place as much as it differs also between current and potential gradient. When one tries to fit a straight line to the cloud of the individual

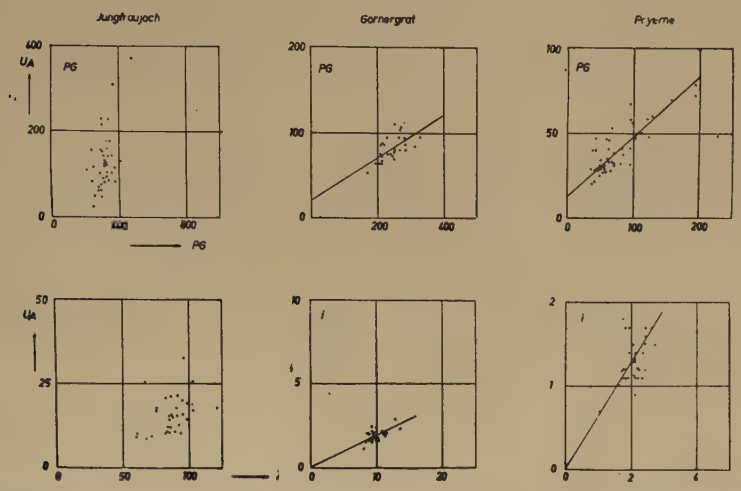


Figure 14. Diurnal mean for the agitation-amplitude U_A of the potential gradient and air-earth current density on undisturbed days in relation to the average value of the concerned factor itself, at Jungfraujoch, Gornergrat and Payerne.

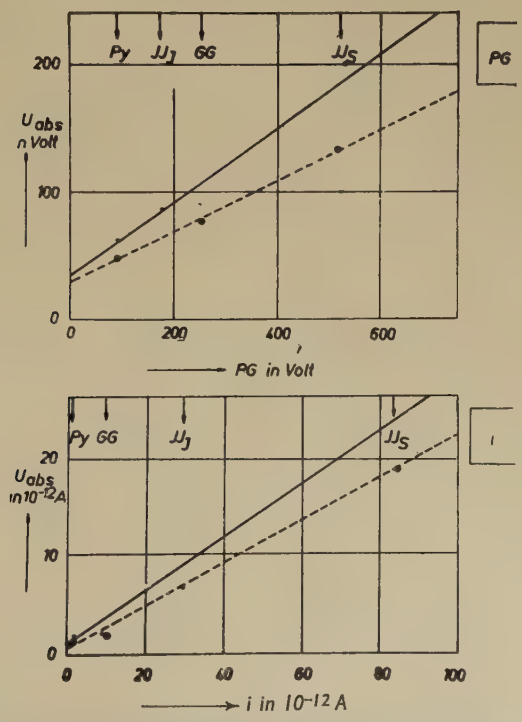


Figure 15. Mean values for the agitation-amplitude of potential gradient (top) and air-earth current density (low) in relation of the concerned element itself at Jungfraujoch-Sphinx (JJ_S) (average for the period from November 1954 to May 1955), Jungfraujoch Institute (JJ_I) (November/December 1954), Gornergrat (GG) (March/May 1955), and Payerne (Py) (December 1954 to June 1955), for the periods without precipitation (points and full-line curve) and for undisturbed days only (circles and broken-line curves).

points, then we get the surprising result shown in Fig. 14. It is shown that for Jungfraujoch and Payerne the curve for air-earth current density passes through the starting point of the coordinates, whereas this is not so in the case of potential gradient. This means, that as a rule the air-earth current density and agitation are proportional but not so with potential gradient.

To test this fact, the total average of the agitation amplitude at the stations is shown in Fig. 15 plotted against the total average value of the element itself during the same period. To supplement the total data obtained at Jungfraujoch, Gornergrat and Payerne, the average values obtained from the two months' recordings at Jungfraujoch Institute have been added*.

Fig. 15 confirms the fact that for air-earth current density the average value of the agitation amplitude is proportional to the magnitude of the current itself. On the other hand for potential gradient, as far as the limited data permit, the agitation amplitude seems to depend partly on two different factors : one which increases proportionally with potential gradient value, and the other which is independent of this value. The proportionality factor between the element value itself and the agitation amplitude is greater for the total material than for the undisturbed days, but is about the same for both potential gradient and air-earth current density.

(*b*) *On the mechanism which causes the agitation*

If we do not consider other factors not interesting here we may conclude that fluctuations in the atmospheric electrical field arise from changes in the condenser potential of the global atmospheric condenser or from changes of the distribution of space charges within this condenser. The first possibility is effective only when the total current circuit of the global atmospheric electricity also shows corresponding changes, and, because of the known relaxation period, this can happen only within the stationary region of atmospheric electrical processes – hence, the first possibility can be ruled out. The agitation can be considered more as a characteristic side-effect of the non-stationary region of atmospheric electricity, which disturbs the immediate surroundings through charge displacement.

If one imagines that a space charge is passing over the observation point at a certain height, then according to the sign of this charge there will be either an increase or decrease in the fluctuations of the records†. When the passing cloud is composed of a dipole with two space – charge packets one above the other, one notices in the course followed by the field records at ground level, either a depression flanked by two peaks or a peak flanked by two depressions‡.

What has been said in the previous section on the agitation in relation to the brightness under a cloud cover (Whitlock and Chalmers (1956) have also given a similar idea) can be considered to be a part of a ' successive individual effect ' as pointed out by Lecolazet (1946) when one makes the supposition that the effect of a charge-generating process in a cloud is proportional to its thickness.

Thus it is clear as to how far the agitation of the current and field at ground level can be traced to causes which originate from the clouds. But since some of this agitation is to be found also on cloudless days, obviously one has to look for other causes which can possibly give rise to them.

It is obvious that this mechanism should be associated with the atmospheric turbulence. Air masses of different origin have their own characteristic atmospheric electrical effects (Israël 1948; 1950a; 1950b; 1950c). The most significant of all is the aerosol content, which because of the different ionization leads to different conductivity and thereby to different space-charge configuration with its own typical behaviour (Israël 1957b, 1957c).

* The Station at Jungfraujoch Institute lies about 130 m lower than the one at Jungfraujoch-Sphinx.

† Examples of this kind show also the effect of smoke clouds which might be passing over the station, as shown by Kelvin (1860) for railway locomotive steam (also observed by Israël 1950a; Mühleisen 1953; Whitlock and Chalmers 1956). A similar case has been observed caused by a volcanic eruption (Hatakeyama 1949).

‡ Examples of this are the field variation with the passage of a thundercloud (see e.g. Simpson and Scrase 1937; Krestan 1941).

When one carries this to the minute air masses of turbulent atmospheric movements, then it is possible to account for the continuous small fluctuations in current and field at ground level.

The mechanism which gives rise to the agitation appears then to be fundamentally the same for cloudy periods as for clear skies. In the case of a clear sky there is obviously no ' stabilizing effect ' (Israël, 1957a; 1957b; 1957c) of the space charge borne on the cloud drops, hence its effect is much less than when the sky is cloudy, as is confirmed by experience (see Section 3 (iii)).

(c) *Causes of agitation*

The properties described above are typical of the atmospheric turbulence processes and especially those of the atmospheric mass exchange processes (austausch), and so it can be said with certainty that the atmospheric agitation is, in general, a direct result of the atmospheric exchange processes. These exchange processes produce turbulent cells which owing to their different aerosol content have different conductivities, and the various space charges thereby produced cause the changes in field and current at the stations over which they pass.

A direct verification of this idea may be possible by determination of the agitation in the conductivity or in the ion content. Indications are that the air conductivity has fluctuations very similar to, and coincident with those of the potential gradient and air-earth current density.

During the preparation of this paper we were informed of a recent study by Brasefield ' Some observations of low-level ion clouds ' (in press 1958). The author investigates the variations of potential gradient with height up to 30 m. He found : ' These measurements indicate that clouds of ions, sometimes positive and sometimes negative, frequently pass overhead at altitudes as low as 10 m and less. The origin of the ion clouds is unknown, but it has been observed that they are more prevalent when the potential gradient is large. The only parameter that appears to be related to the prevalence of the ion clouds is the electric field itself. When the potential gradient is high (e.g., 250 volts per metre), the number of low-level ion clouds of either sign which pass per hour is usually high (6 to 8 per hour) whereas when the potential gradient is 100 volts per metre or less, less than one low-level ion cloud passes per hour.'

Supposing that there are no differences due to the measuring method, we may assume the ion-clouds postulated by Brasefield are identical with turbulent cells of different space charge postulated by us.

5. Conclusions

The experiences on atmospheric electrical agitation which are given above serve on the one hand, to indicate the close coupling between meteorological and atmospheric electrical phenomena, and on the other hand, they lead us closer to the idea that we can use this relation obtained from the agitation as much in meteorology as in atmospheric electricity.

The first task is to investigate the exchange process in the light of the experience gained in work on atmospheric electricity; this is only a continuation of the earlier investigations done by the author which follow the changes in field and current at different altitudes, and from their actual variation to get a better insight into the exchange processes at higher altitudes. Another direct-exchange experiment can be thought of, which consists in the study of the behaviour of the atmospheric electrical factors, when the air is marked by radioactive materials (see the work of this kind done in Russia during the last four years : e.g., Milin 1955).

The separation of global and local influences by choosing the ' undisturbed days,' as usually attempted, is insufficient since it is effective only where the local influences are small compared with the global influences. This is the case only over the oceans or in high mountain but *not* in flat-land situations. In the field of atmospheric electricity a study of the agitation looks very promising as a help in this problem of separation.

Many local influences are to be seen in the agitation, and it is suggested that one should look for a correlation between these factors and the degree of the agitation of potential gradient and air-earth current density.

As the material described here is not sufficient for a more thorough investigation it is difficult to give more details. However, it is hoped that the way has been pointed to showing the crucial problem of separating the two influences in the sphere of atmospheric electricity. It would be of particular interest if work of a similar nature could be done in geographically different situations on land and also in mid-ocean.

References

Bidlingmeyer, C.	1913	*Veröffentl. d. Kaiserl. Obs. in Wilhelmshafen.* Halbj. 1911, Berlin 1913.
Chalmers, J. A.	1957	*Atmospheric electricity*, Pergamon Press, London.
Courvoisier, P.	1951	*Experientia*, **7**, pp. 241-254.
Dolezalek, H. and Fries, G.	1956	*Tech. Note* **7**, Contract AF 61 (514)-640.
Hatakeyama, H.	1949	*J. Geomagn. Geoelectr.*, Japan, **1**, No. 2, pp. 48-51.
Hess, V. F.	1926	*Die elektrische Leitfähigkeit der Atmosphäre und ihre Ursachen.*, Verlag Vieweg, Braunschweig.
Israël, H.	1939	*Gerl. Beitr. Geophys.*, **55**, pp. 314-333.
	1943	*Met. Zeitschr.*, **60**, pp. 174-177.
	1948	*Met. Rund.*, **1**, pp. 200-204.
	1950a	*Ann. Met.*, **3**, pp. 329-332.
	1950b	*J. Atmos. Terr. Phys.*, **1**, pp. 26-31.
	1950c	*Arch. Met., Geophys. Biokl.* (A), **3**, pp. 1-16.
	1957a	*Geofisica pura e applicata*, **36**, pp. 182-200.
	1957b	*Tech. Note* **11**, Contract AF 61 (514)-640, Nov. 1956.
	1957c	(*Alps-Project* 1954-57) Final Rep., Contract AF 61 (514)-640.
	1958	*La Météorologie* (in press).
Kelvin, Lord	1860	*Phil. Mag.*, **19** (and B.A. Paper on Electrostat. and Magnetism), pp. 316-320.
Koenigsfeld, L.	1955	Proc. Conf. Atmos. Electr., *Geophys. Res. Pap.*, **42** (AFCRC), pp. 21-26.
Krestan, M.	1941	*Gerl. Beitr. Geophys.*, **57**, pp. 334-356.
Kuhn, H.	1933	*Zeitschr. Geophys.*, **9**, pp. 238-252.
Lecolazet, R.	1946	*La Météorologie*, Oct.-Dec., pp. 429-433.
Milin, W. B.	1955	*U.S.S.R. Glavnoe Upravlenie Gicromet. Sluzhby*, Informat. Sbornik, **1**, pp. 28-36.
Mühleisen, R.	1953	*Zeitschr. Geophys.*, **19**, pp. 142-160.
Nordmann, C.	1911	*Ann. du Bur. des Long*, **8**, p. 3-60, Paris.
Rossmann, F.	1950	*Ber. Deutsch. Wetterd.* US-Zone, **3**, No. 15.
Schmidt, W.	1909	*Wien. Ber.* (IIa), **118**, pp. 885-897.
	1913	*Ibid.*, **122**, pp. 835-911.
Simpson, G. C.	1949	*Geophys. Mem. Met. Off.*, **10**, No. 84.
Simpson, G. C. and Scrase, F. J.	1937	*Proc. Roy. Soc.* A, **161**, pp. 309-352.
Whitlock, W. S. and Chalmers, J. A.	1956	*Quart. J. R. Met. Soc.*, **82**, pp. 325-336.

551.578.7 : 551.501.81

Patterns of hailstorms in Alberta

By R. H. DOUGLAS and W. HITSCHFELD
*Meteorological Service of Canada** *McGill University*†

(Manuscript received 24 July 1958; in revised form 7 Nov. 1958)

Summary

Between 14 June and 20 September 1957, nearly 3,200 surface reports of hail were collected in a 15,000 sq. mile area. Information included location, time and duration of hail and the largest size of hail encountered. Continuous radar records of the storm pattern were taken in a manner suitable for the preparation of constant-altitude radar weather (CAPI) maps, to a range of 100 miles.

At times, hail fell soon after a first echo (possibly within 10 min), sometimes no hail occurred for as long as an hour or two. Usually, the hail fell in short bursts (lasting up to 20 min and spread over a few miles) which would appear to be the products of single storm cells. But on several occasions (notably on 23 July) hail fell continuously for 1½ hours and, without evidence of a multiplicity of cells, covered a strip 30 mi long. Sustained hail came from storms whose echo tops were steady close to 30 kft above surface. Hail associated with echoes of lesser height was not continuous and tended to fall in bursts which sometimes followed closely on the merger of previously separate storms. A study of 83 independent echoes showed that the probability of hail from a given storm was clearly related to maximum height of the storm, rising from zero for storms 10 kft or lower, to 100 per cent for storms 38 kft or more in height. A similar correlation holds between echo height of a hailstorm at any instant and hail arrival at the ground some minutes later.

1. Introduction

Reports of hail from observers on the ground provide invaluable detail about the surface pattern of the hail but tell little about the structure aloft of the parent storms. Radar, on the other hand, reveals the history and structure of the three-dimensional precipitation patterns (but not of the clouds). Its greatest weakness for our purpose is its failure to distinguish between rain and hail‡. Our Alberta study combined hail reports with radar observations. Surface hail was located in space within half-a-mile and in time within 5-10 min; the maximum stone size was reported subjectively to a reasonable first approximation. The storm echoes were located by radar with a precision better than one mile; the development in time was known with a precision better than 2 min. Photography yielded records of samples of cloud formations and their variations, but no quantitative cloud measurements were made.

The project area comprised 15,000 sq. mi of notorious hail country in central Alberta (Fig. 1)§, and consisted chiefly of flat prairie land, except for a few rolling hills and the rather steep coulees cut by two or three meandering rivers. The elevation in this area rises uniformly from 2,500 ft above m.s.l. in the northeast through 3,000 ft at Penhold, where the radar was located, to 4,000 ft in the southwest. To the west and southwest, towards the foothills of the Rocky Mountains, the population decreases fairly rapidly, becoming quite sparse beyond about 30 mi from Penhold. Thus there is a severe limitation to the distance westward from which hail reports could be expected. Since many storms moved into the project area from that direction, data on first hail must often be incomplete.

‡ Harrison (1956) reports the identification of hail by means of characteristic fingers and scallops on the edges of intense storm echoes. No equally distinctive pattern has been found in our records, which however have not been examined in detail at comparably short range.

§ In roughly 3,000 sq. mi in this area the Water Resources Development Corporation Modification Corporation of Calgary was attempting a commercial hail suppression programme, whose operations were quite independent of ours. No special evaluation of their work was undertaken, but there was no indication in the data we collected that observations in this area should be treated differently in any way from those originating elsewhere.

* On assignment to the Stormy Weather Group, McGill University. Publication is with permission of the Director of the Meteorological Service.

† The processing and analysis of the data at McGill University were supported by the Research Council of Alberta.

The data reported here were gathered under the sponsorship of the Research Council of Alberta with the assistance of the Meteorological Service and the National Research Council; R. H. Douglas was scientific director.

Figure 1. Site of the Alberta Hail Research Project. The circle of 100-mi radius shows the region covered by the radar at Penhold. The hail-reporting area is shown in solid outline, and the area under contract for hail suppression in dashed outline.

More than 15,000 householders in the project area were supplied with printed postcards designed to minimize the effort required to complete the hail report. Information requested included the size of the largest stones observed, according to an arbitrary size-scale : shot (3 mm diameter), pea (6 mm), grape (20 mm), walnut (30 mm), golfball (50 mm), and larger. It was emphasized that all hail occurrences, no matter how small and harmless the stones, were to be reported. In addition, the records of the hail insurance companies permitted us occasionally to narrow down spatial gaps in the reports. Hail reports were received on 77 out of 108 days in the period 4 June to 20 September; in all, the reports numbered 3183 (see Fig. 2). The largest hailstones occurred most frequently near the end of July, remained fairly common in early August, but declined rapidly thereafter. When presented on a monthly basis this seasonal trend agrees quite well with the observations made by Beckwith (1956) in Denver, Colorado, in 1949 to 1955.

All but the extreme southern portion of the project area was under the surveillance of the weather radar at Penhold. Radars of ordinary power do not detect the clouds themselves but reveal only the precipitation within them. To study the structure and development of the precipitation in such clouds the radar beam should be narrow compared to the storm dimensions. Our radar, specially assembled by Decca Radar (Canada) Limited, provided a beam $\frac{3}{4}°$ in height and 4° in horizontal extent. The radar operated at 3·2 cm wavelength. The pulse length was 5 microsec; pulses were transmitted every 2·5 millisec. Peak power was 30 kw, which was adequate for the detection of the storms to be studied. A study of several days' records indicates that the number of storms per unit area of surface which was detected was independent of range up to 40 mi. Beyond this range, where the beam is less likely to be filled by target precipitation, the number of storms per unit area detected, varied approximately inversely as the cube of the range. No storm echoes were received from farther than 80 mi. The radar was equipped with an antenna control device which with the good resolution in height made three-dimensional analysis of the storms possible.

In the scanning programme of the radar, the antenna rotated continuously and was elevated automatically by one-half degree each revolution. A time-exposure photograph of the display was taken on a frame of 16 mm film covering two successive revolutions in azimuth. Thus a series of (approximately) one-degree slices through the storm was

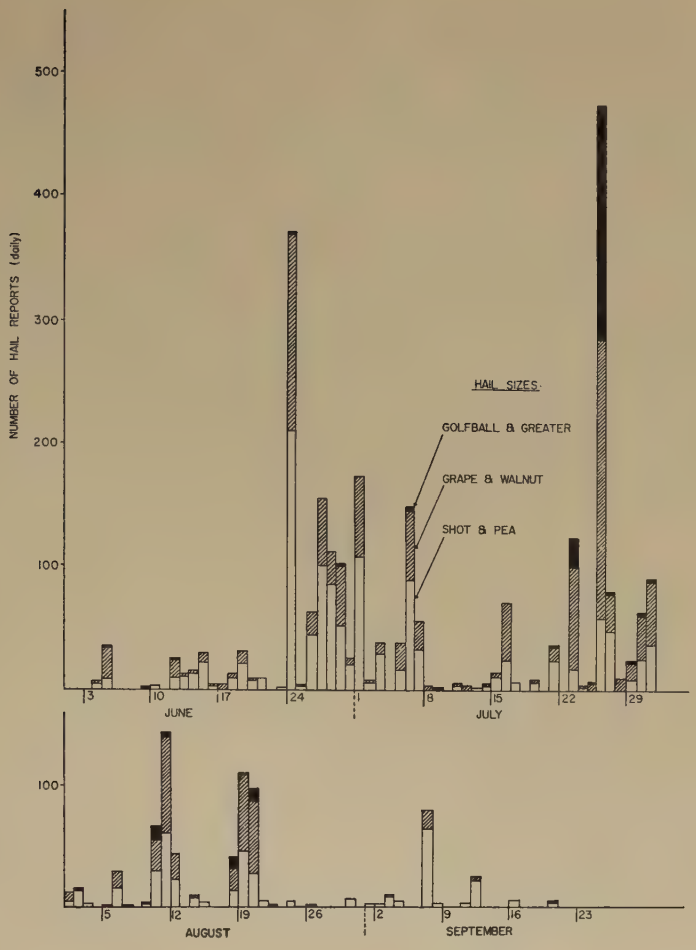


Figure 2. Number of hail reports received and maximum stone-sizes for each day during the project, summer 1957. The most destructive day with 479 hail reports was 26 July, for which radar records are not complete.

recorded on successive frames of film, the slices cutting through the storm at progressively higher levels. The antenna rotated at 20 r.p.m. When it reached a maximum elevation of 20 degrees, it returned to the horizontal to repeat the cycle. A complete scanning cycle took about 2·2 min. It is believed that this way of scanning the atmosphere is most economical in time, and because it avoids the frequent starts and stops of the antenna in a succession of vertical scans, it is both faster and easier on the radar mechanisms. Another advantage of the quasi-spiral climb of the antenna beam is that it permits the automatic synthesis of *constant altitude* (CAPI)* displays, which show precipitation maps at any desired altitude above ground. The equipment used for such synthesis was specially designed (Hitschfeld (Ed.) 1959).

2. Observations

16 July was one of five days selected for detailed analysis because of the number and variety of the individual storms, most of which spent the greater part of their histories under radar surveillance and within the hail-reporting network. From a deep low over western Hudson Bay, a front trailed across the southeastern corner of Alberta. Surface circulation over the project area was weak northwesterly, but winds aloft backed into the

* The first application of CAPI displays was described by Langleben (1956).

southwest ahead of a 500 mb trough; this trough extended between closed lows over northeastern Alberta and southern British Columbia. Upper-air soundings for this day (and the others studied in detail) are shown in Fig. 3. Fig. 4 shows in outline the storm envelopes, the strip-like areas swept out by the radar echoes (derived from the CAPI plan-pattern synthesized for height 5 kft). Such envelopes do not indicate the actual shapes

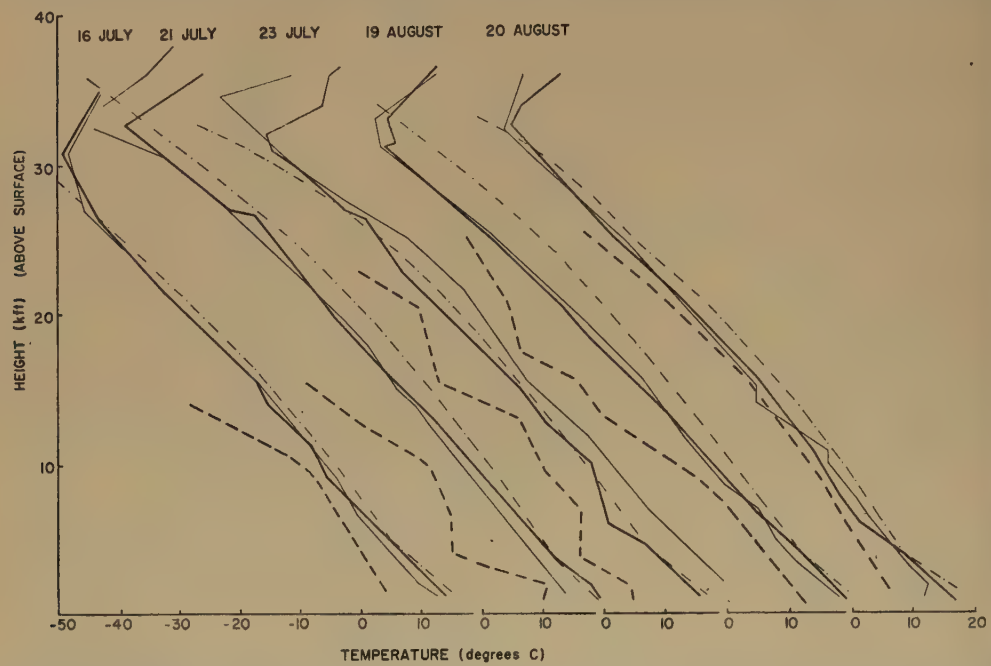


Figure 3. Temperatures aloft over Edmonton at 0500 MST (full thin curves) and at 1700 (full heavy curves) and dew points at 1700 (dashed curves) on the 5 days indicated. The dry adiabats below cloud-base and the cloud adiabats above are also shown (dashed-dotted curves). (Successive temperature scales are displaced by 20°C; heights are with respect to Penhold, elev. 3,000 ft).

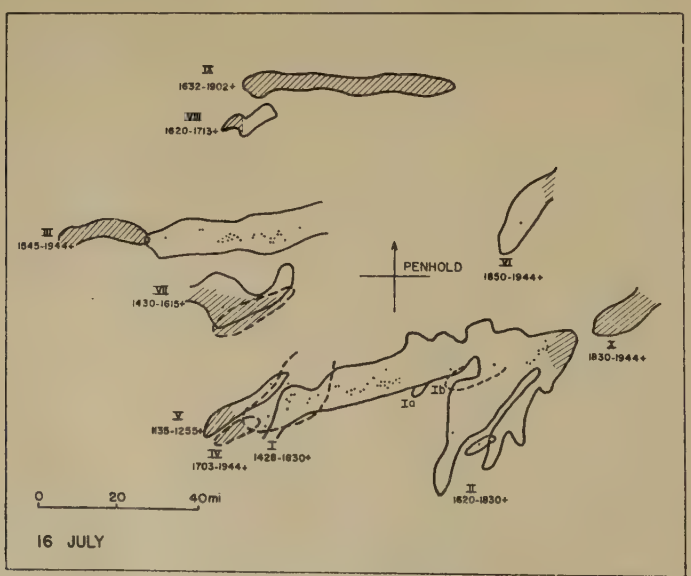


Figure 4. Hail observations and envelopes of storms. The envelopes show the areas swept out by the radar echoes at 5 kft. Areas which are swept out by these storms, but from which hail reports were not solicited, are shown hatched.

of the storms or their radar echoes at any given instant, which are roughly circular, but the areas over which the storm passed and upon which precipitation presumably fell. All storms throughout the summer moved in an easterly direction; hence the motion of any storm in its envelope should be visualized as proceeding from left to right. The dots on the figure locate the day's 71 hail reports. Of these 21 were of pea, 42 of grape, and 8 of walnut-sized stones. Nearly all the hail that was reported occurred along two well-defined strips.

Storm I was well developed, with echo top at 30 kft (Fig. 5) when first observed. First reported hail fell at 1440 MST and was of grape size, as was most of the subsequent hail. At about that time the top of the storm subsided to 25 kft but surged up to 32 kft by 1520, then gradually subsided during the next two hours. The most dense hail-fall occurred during this gradual subsidence. The densest portion of the hail strip was somewhat closer to the southern edge of the storm than to the northern, and lagged by 20 min the time of the highest echo top. During the 90 min following this burst of activity, there were only two further reports of minor hail attributable to this storm. Each of these two occurrences followed closely on the merger of smaller echoes I*a* and I*b* with Storm I. In the second case it appeared as if the secondary echo grew rapidly and became the main one, as is suggested in Fig. 5.

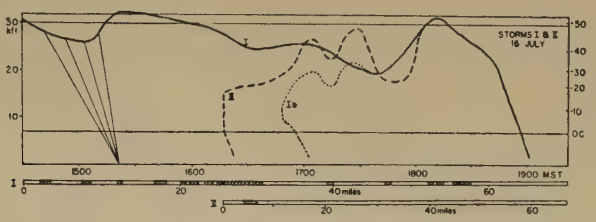


Figure 5

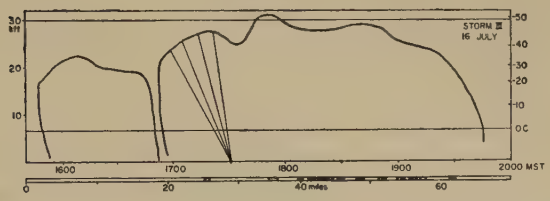


Figure 6

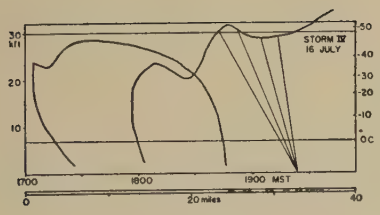


Figure 7

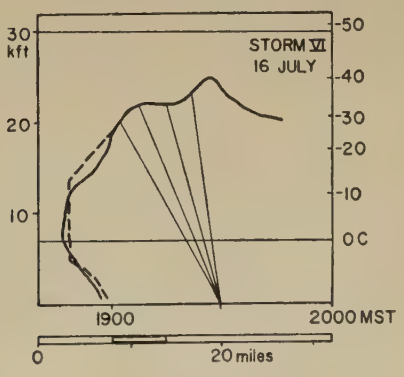


Figure 8

Figures 5, 6, 7 and 8. Echo height as a function of time. Hail reports are located according to their times along the auxiliary distance scales at the bottom. (Open rectangles for small, hatched for medium-size hail). Temperatures aloft are shown on the right-hand ordinate, and the heights (above surface) of the melting level and of the tropopause are indicated. Fan-like straight lines suggest heights and times at which hail formed in the storm to arrive at ground at time of fan apex. (These lines are drawn for particle fall speeds of 2, 4, 8 and 16 m sec^{-1}).

Meanwhile a cluster of smaller storms (II) appeared to the southeast about 1630 MST These moved northeastward, combining one with another, and by 1800 joined Storm I. The top of the dominant member II*a* of the complex II had stayed at modest heights and by itself had produced but one isolated occurrence of small (pea-sized) hail. But as II*a* approached Storm I and presumably entered its local vertical wind field, rapid vertical development resulted, which was accompanied by a fresh hail-fall which lasted altogether for about 50 min, seven reports of grape-sized stones being received from a relatively small area. Both Storms I and II, showed marked vertical development and revival of hail activity associated with the confluence or junction of several storms. Their apparent convergence, at an angle of 50°, while still separated by as much as 40 mi, was also noteworthy. Storm motion usually reflects the atmospheric circulation only in a general way, and the failure of neighbouring storms to move in parallel paths can often be ascribed to an interaction between them. But the suggestion of an interaction over a distance of 40 mi would surely need further consideration.

Storm III was first detected aloft (Fig. 6). After achieving a maximum height of about 22 kft it subsided first gradually for one half-hour, then rapidly; by 1652 it appeared only as a scattering of weak echoes. Between 1652 and 1654 a remarkable redevelopment took place, leading to a large intense echo of considerable duration, which eventually reached up to 30 kft, the height of the tropopause. The top remained near 30 kft for one hour before subsiding gradually. The largest hail (walnut), and the greatest durations of hail-fall, occurred over a half-hour period beginning shortly after the achievement of the maximum echo height. Echo decay and rapid development of the kind noted in Storm III was observed on a few other occasions. The occurrence of the first hail, within minutes of a new echo, has also been noted elsewhere in this study.

The hail produced by Storm IV fell at the extreme southwestern end of the hail strip laid down several hours earlier by storm I. Storm IV consisted of two individuals, the first of which appeared at 60 mi range outside the reporting area. Detected first between 19 and 24 kft, the echo reached a maximum height of 28 kft (Fig. 7), and had a lifetime of nearly 2 hr; it apparently shed no hail. During its last half-hour, a parallel and separate development occurred some 10 mi away, appearing first between 11 and 18 kft. Hail from this was reported within 33 min, while the storm required 50 min to develop to a height of 32 kft. Hail was generally of grape size, but two observations of walnut-sized stones were made. Most of the hail was associated with echo tops of 30 kft, not with the greater heights reached subsequently.

Of the remaining storms within radar range on this day only Storm VI (Fig. 8) was known to produce hail, and that within 10 min of its first detection, while the echo top was rising (at about 2·4 m sec^{-1}) to a maximum height of 25 kft. It may well have produced

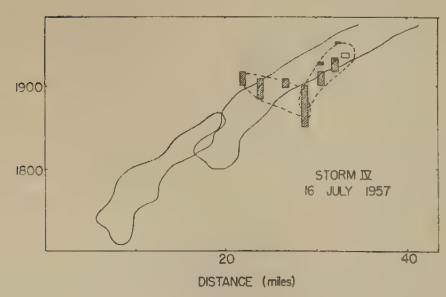


Figure 9

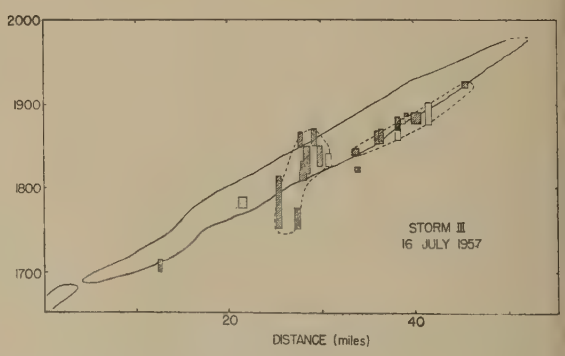


Figure 10

Figures 9 and 10. Times of hail fall and of arrival of the forward and rear edges of the 5 kft echo as functions of distance along the hail strip measured from an arbitrary origin. Hail sizes are indicated as follows : small (shot and pea), open rectangle; medium (grape and walnut), hatched; large (golfball and greater), solid.

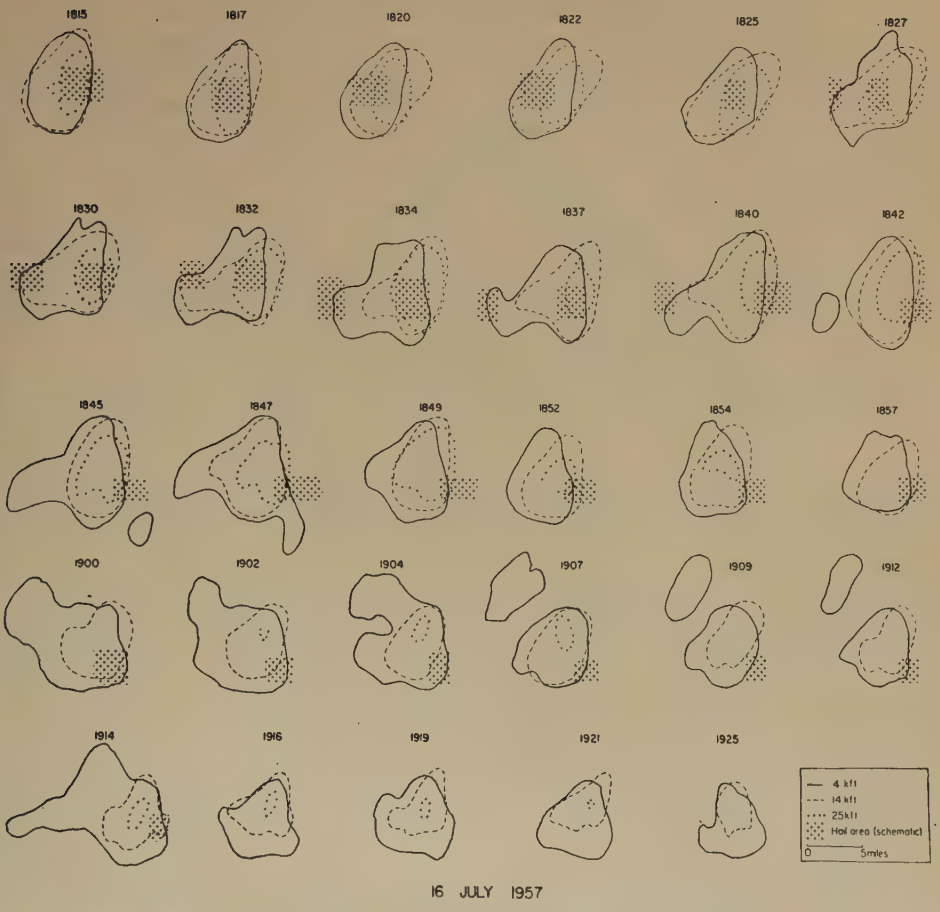


Figure 11. Horizontal sections through radar echoes of Storm III of 16 July at heights 4, 14, and 25 kft during a period of widespread but diminishing hail activity. Areas on the ground where hail was reported are indicated schematically. Location of echoes was 30 mi due west of the radar.

hail later on, but since it moved out of our network no further information became available. Storm IX probably also produced hail, but the part of the countryside over which it moved did not become part of our reporting network till later in July.

To study storm motion in detail, and to tie hail and storm echoes together more closely, ' x-t ' graphs were prepared for most of the storms. The times of arrival of the forward and rear edges of the storm echo at 5 kft were plotted against distance along the hail strip measured from an arbitrary origin. An example (for Storm IV) is Fig. 9. The two components of this storm are seen to exhibit a similar change in pace : during their first half-hours, each echo moved quite slowly, but subsequently accelerated. Although the second storm developed some 10 mi ahead of the first, it was overtaken during its early slow-moving stage by the first. While no junction occurred, the path of the second storm is an exact continuation of the path of the first. We also entered the hail information on the ' x-t ' diagrams. In interpreting the relation of the hail to the echo in Fig. 9, one could take the stand that the reported hail times are likely to be in error, and that in reality the rectangles representing the hail fall should all be moved up or down (not horizontally : the people reporting know their location), until they fall well within the storm envelope. More constructively, taking all reported times at face value would open up the possibility that within one storm two hail cells were at work, with one playing along the leading edge, the other at the same time cutting back through the storm. This possibility is suggested, in somewhat exaggerated fashion, by the dotted outlines around the hail rectangles. In

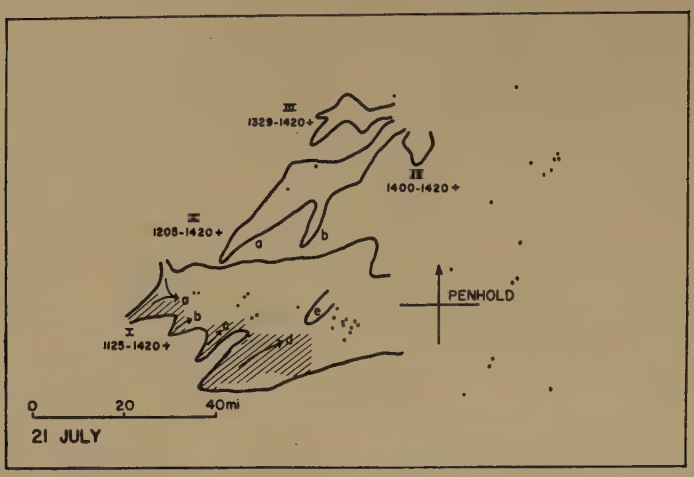


Figure 12. Hail observations and areas swept out by the radar echoes at 5 kft. Areas which were swept out by these storms, but from which hail reports were not solicited, are shown hatched.

sketching these outlines no allowance was made for the probable error in the vertical position of the largest rectangle in the figure. The same question of the accuracy of the reported times of hail fall arises in Fig. 10, the ' x-t ' graph for Storm III. On the one hand, two observations of hail at 1730 stand out from the others, and their reported times seem rather premature. If they are correct, they would imply the throwing of hail ahead of the storm by as much as 10 mi, which in view of the prevailing wind field would seem impossible. On the other hand, the reports do suggest that the hail-fall may again have resulted from two separate hail generating regions in the storm, as is suggested, again in exaggerated fashion, by the dashed borders drawn around the hail reports. The reported times after 1815 are unexceptionable, however, and we have used them to construct idealized areas of hail fall which were fitted to sketches of the radar echo drawn at heights 4, 14 and 25 kft (Fig. 11). The resulting patterns indicate that the hail was initially associated with the summit of the storm, then advanced to the forward edge, at times preceding it by as much as 4 mi. At 1827 a rearward extension of the echo was accompanied by a separate shaft of hail, which persisted until the dissipation of this extension at 1842.

On 21 July nearly half the hail was laid down along a fairly narrow strip by a series of independent storms; another 20 sporadic hail occurrences came from numerous smaller echoes. Alberta lay in an elongated surface trough. During the morning a trough of warmer air aloft moved eastward across the province. Low-level winds were generally southwesterly, but above 10 kft, westerly winds predominated. Between 0500 and 1700 the air below 500 mb (15 kft) warmed and became less stable, particularly below 3 kft (Fig. 3). The most notable change was in the moisture, which decreased during the day at all but the lowest levels. In the early morning a moist layer extended up to 15 kft, but 12 hr later this ' moist ' layer had dried considerably and extended only to 10 kft. The convective condensation level in the afternoon was at 4 kft.

The first echoes *a* of the line of Storms I – two in number – were already detectable when observations began at 1125 (see Fig. 12). During the next three hours, three substantial echoes *b*, *c* and *d* (and a smaller echo *e*) formed to the south and one after the other moved up to a W-E line, west of the radar, where they reached their maximum development, produced some hail, and then dissipated. The remnants of each of these stayed on along the line, merged with the next storm, thus giving (in Fig. 12) the impression of a single ever-widening storm moving eastward, into which secondaries streamed from the south and west. The line I thus appeared to be travelling eastward, with each shower moving NNE-ward, which was also the direction of most of the more isolated showers that day. The appearance of hail along a fairly well-defined W-E strip is striking, suggesting that hail formation commenced in each shower after it had entered an environment

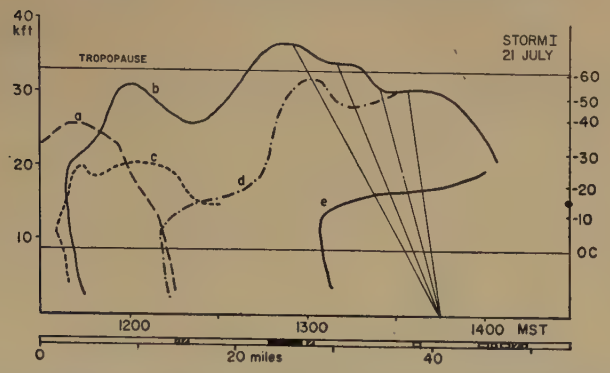


Figure 13. Height/time diagram for Storm I, 21 July. Hail occurrences are indicated at bottom : small hail, open; medium size, hatched; large, solid. Fan-like lines drawn for particle fall speeds of 2, 4, 8 and 16 m sec^{-1}.

modified by the remnants of the earlier storms. The hail reports both in time and location seem to fall into four discrete groups related to *a* and *b*, to *c* to *d* and to *e*, respectively (see also Fig. 13). Some 20 to 30 min after the last onset of hail the line I passed over the radar site, accompanied by gusts up to 70 mi hr^{-1} which disrupted radar observations.

The major Storm I of 23 July (Fig. 14) passed through an area which had a long-term record of destructive hail, and resulted in extensive crop damage along a 40 mi strip. During the day an elongated low crossed the province, the centre moving NE-ward during the forenoon. A cold front followed about mid-day, and in the afternoon the region lay in a NWly current behind the system. Winds aloft had been S to SWly in the early morning but by 1700 those below 12 kft had veered into the W and NW. Marked cooling occurred over Edmonton between 0500 and 1700 at all levels below the tropopause, and the freezing level fell from 11 to 8 kft (Fig. 3). At 1700 the convective condensation level was just below 5 kft.

All but a dozen of the 126 hail reports came from Storm I, and 31 of these were of stones the size of golfballs or greater. The first traces of this storm were already detectable by 1300 MST, 80 mi to the NW, with tops near the modest height of 15 kft (Fig. 15). Big development in both height and intensity took place only after more than one hour. During the next hour the top ascended in steps to the tropopause at 32 kft, and it remained at this height for nearly another hour before descending, at first gradually, then very rapidly. Hail was plentiful from 1415 to the very end of the storm, and though of assorted sizes, the maximum sizes reported showed a definite trend, which is indicated in the figure.

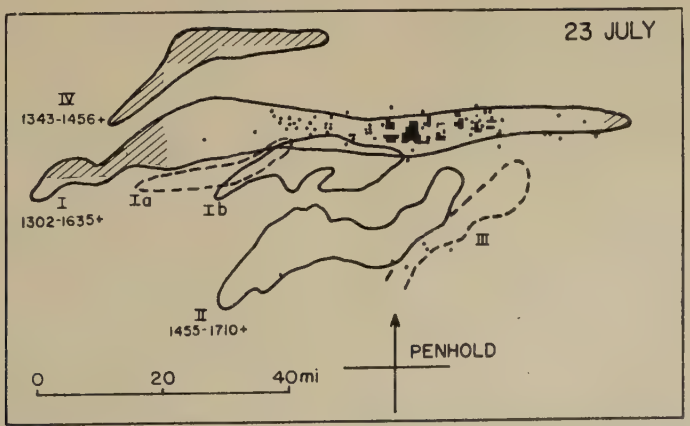


Figure 14. Hail observations and areas swept out by radar echoes at 5 kft. Areas which were swept out by these storms, but from which hail reports were not solicited, are shown hatched.

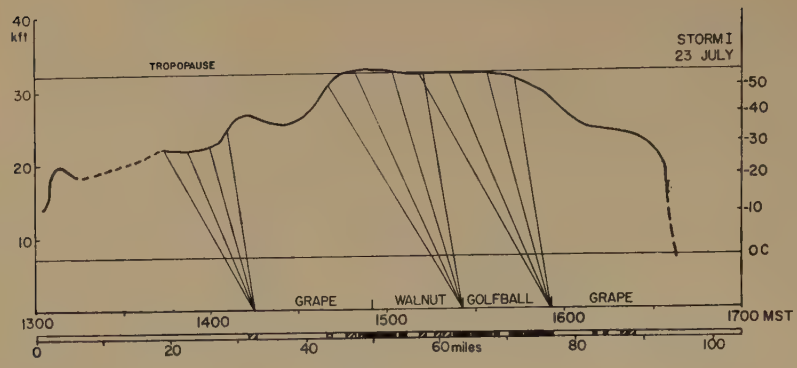


Figure 15

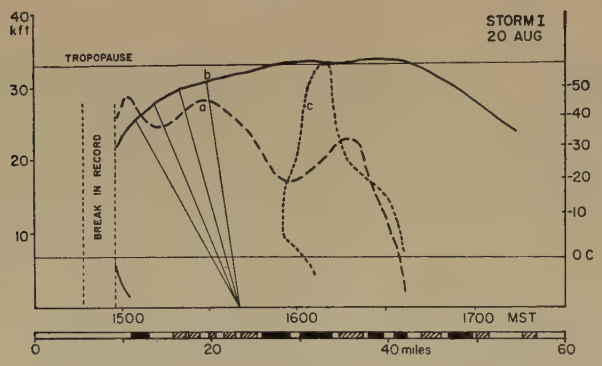


Figure 16

Figures 15 and 16. Height/time diagrams for Storm I, 23 July, and for Storm I, 20 August. Hail occurrences are indicated at bottom : small hail, open; medium size, hatched; large, solid. Fan-like lines drawn for particle fall speeds of 2, 4, 8 and 16 m sec^{-1}.

When the trajectories of these stones are taken into account, a possibly highly significant correlation between hail size and echo height appears : grape-sized hail is associated with the lower echoes, walnut with the echo upsurge, and golfball with the maximum heights. This storm appears to have been an almost continuous producer of hail over a distance of some 40 mi, over a time of about 1½ hr. This feature distinguishes it from the storms of 16 and 21 July, which hailed intermittently and whose hail production apparently tended to be associated with the junction of secondary storms, or with separate storms in one complex. In the immediate vicinity of Storm I, echoes designated I*a* and I*b* were noted. These reached maximum heights of 25 kft or less, and lasted for about one hour each. Though they came rather close to the hail strip they were probably not responsible for any of it. Storm II also in the vicinity reached a maximum height of 25 kft, and though it lasted for more than two hours and was quite steady at its maximum height, gave rise to but two hail reports.

Another example of a major hailstorm is that of Storm I of 20 August. On this day a flat pressure field lay over the area without moving. Low-level circulation was slack, but above 10 kft it was southwesterly, with speeds rising up to 77 kt. A total of 96 hail reports were received, of which nine were of very large stones (golfball or greater). Storm I (Figs. 16 and 17) was somewhat reminiscent of our previous example (Fig. 14), except that it started as two separate echoes *a* and *b*, and included a later subsidiary cell *c* also. At 1458 two echoes (*a* and *b*) were visible to the SW, both having developed some time in the preceding 12 min for which no radar record is available. The nearer one *b*, moved but little at first, while *a* advanced steadily for an hour, dissipating to the rear of *b*. Though *a* reached considerable height no hail was reported from it. On the other hand,

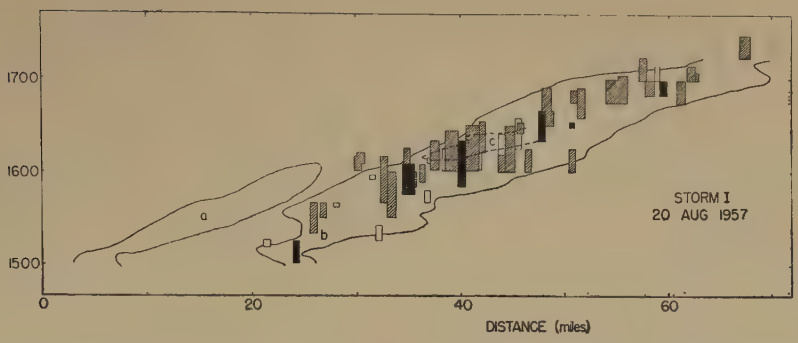


Figure 17. Times of hail fall and of the arrival of the forward and rear edges of the 5 kft echo as functions of distance along the hail strip measured from an arbitrary origin. Hail sizes are indicated as follows : small (shot and pea), open rectangle; medium (grape and walnut), hatched; large (golfball and greater), solid.

b was already producing hail when first observed, within less than 17 min of the first formation of detectable echo. The first hail observation at 1500 indicated stones as large as golfballs ! During the time that the echo top was gradually rising, hail fall was sporadic although the stones were large. Once the top achieved its maximum height of 32 kft (at 1540), the hail became almost continuous, and remained so until the echo top began to subside at 1630. The continuous hail included all sizes from pea to golfball. The subsidence of the echo was accompanied by sporadic but large hail. Fig. 17, the ' x-t ' graph of the storm, shows the relative timing of hail and echo. It is included here to illustrate the mutual consistency of the times reported, and their good agreement with the times of the echo.

3. The observations in summary

The storms of the days we analysed seem to fall into two groups : on 16 and 23 July for instance, the storms swept out relatively narrow tracks, and the hail from them covered well-defined swaths. On other days, e.g., 21 July, the storms or the alignments they formed were more extensive, covering much broader tracks, and the hail tended to be more widely scattered. Such hail swaths as there were, were patchy and relatively diffuse. Hail size (and damage) was not significantly different on these two groups of days.

On the days of narrow storm tracks, the echo of a well-developed storm might be 10 mi in diameter, last for 3 hr, and cover some 80 mi in a NE to E direction. The hail would typically be associated with the echo top reaching to, and remaining steady near, 30 kft. The hail swath might start some 30 mi from the beginning of the storm, and extend for 40 mi; its width might be 5 mi. In hailstorms whose echo tops hovered near 30 kft, or were steady well below that level, the hail was generally patchy. Often, short bursts of hail followed the convergence and merger of two separate storms. On the other group of days, the storms at any instant were bigger, or they tended to form big complexes, up to 30 mi in diameter, and though they persisted again for up to 3 hr and covered long tracks, the echo envelopes were nearly as wide as they were long. The appearance of hail again related in a general way to echo height, but the connexion was less clear-cut. One possible reason for the difficulty in establishing this connexion is that new cells were continually developing at the edges without completely merging; this prevents the assignment of an echo height to a given hail fall. When hail swaths did occur, they sometimes extended through the whole width of the echo track. But even when the hail was heavy, the swaths were not coherent.

There were no obvious differences among the soundings (shown in Fig. 3) of the days just reviewed. The soundings (taken at 1700 MST over Edmonton, about 100 mi from the centre of our area) generally indicated relatively dry air aloft, but usually failed to exhibit the moist lower layers, typically surrounding severe storms in the American Midwest. The differences between the soundings on the days with severe hail and on days without any hail were also slight; so slight, in fact, that we found it impossible

to verify the Fawbush and Miller (1953) and Foster and Bates (1956) techniques for relating hail to large positive areas on the tephigrams between the environmental dry-bulb curve and the parcel adiabat.

TABLE 1. HEIGHTS, TEMPERATURES, AND FIRST-ECHO ADIABATIC LIQUID-WATER CONTENTS, OF STORMS ON 5 DAYS

	16 July	21 July	23 July	19 August	20 August
Tropopause height (kft)	31	33	31	31	33
Echo top, average height (kft)					
Hail-producing storms	29	38	29	35	29
Storms not producing hail	25	—	22	—	23
Echo top, average temperature (°C)					
Hail-producing storms (*T* denotes tropopause temperature)	− 47	− 59 (*T*)	− 47	− 56 (*T*)	− 48
Storms not producing hail	− 40	—	− 30	—	− 32
First echoes					
Average height of top (kft)	18	17	17	21	17
Temperature (°C)	− 23	− 13	− 14	− 29	− 18
Liquid-water content (gm m^{-3})	3·5	4·5	3	4·2	3·7
Average height of base (kft)	10	10	8	9	10
Temperature (°C)	− 6	− 2	− 1	0	− 5
Liquid-water content (gm m^{-3})	1·8	2·8	1·5	2·1	2·8

It is instructive to examine the very beginnings and subsequent developments of the storms. On five days which were examined in full detail there were 39 echoes whose initial top and base heights, and the maximum heights they reached in their maturity, are shown in Fig. 18. Whether or not hail was associated with a given storm is also indicated, where possible. Averages for each day are set out in Table 1, together with temperatures and calculated liquid–water contents. To judge by this Figure and the Table, most hail-producing storms tended to reach close to, or break through, the tropopause; the storms without hail invariably fell far short of this level (but see Fig. 19). The calculated adiabatic liquid-water contents and the temperatures prevalent at first-echo heights suggest

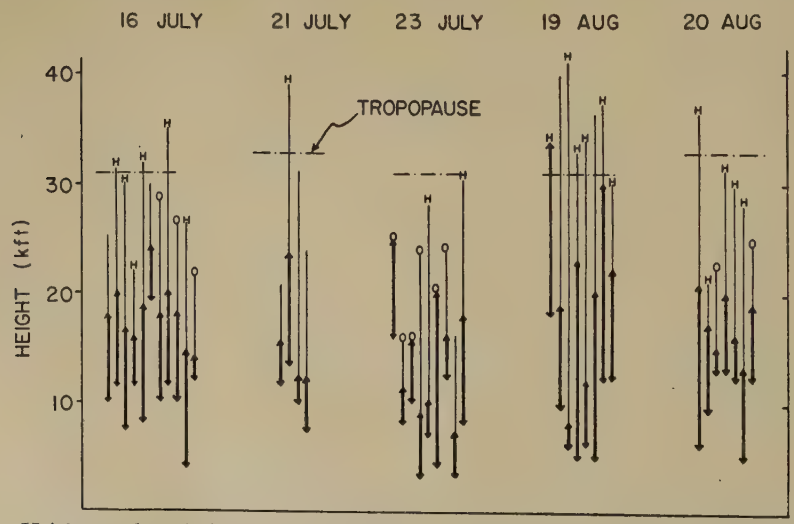


Figure 18. Heights, and vertical extents of first echoes (thick lines), and of maximum echo tops (thin lines) for the storms on five hail days for which first-echo information was available. Hail-producing echoes, and non-hail-producing echoes, are indicated by H and O respectively; echoes for which no hail data were available are unmarked.

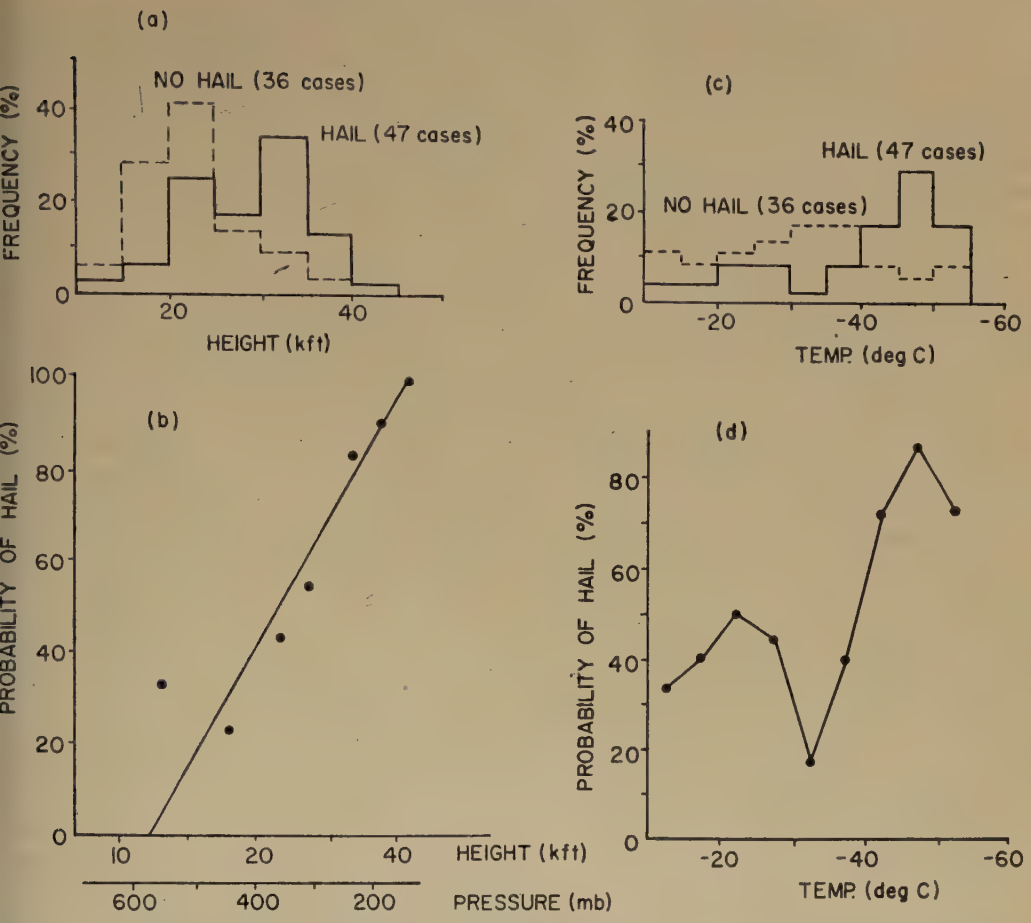


Figure 19. Distribution with (a) height and (c) temperature of echo top of 47 storms which produced hail and of 36 which failed to do so. Probability of hail as a function of (b) height (logarithmic) and pressure, and (d) temperature. (Probability = number of hailstorms of given maximum height or echo top temperature divided by total number of storms of that height or temperature.)

that precipitation growth would *commence* by two processes : the cloud densities were great enough to permit the initiation of precipitation by coalescence (East 1957), the temperatures low enough to provide a large number of crystals for further growth by diffusion. But the variations in these temperatures and liquid-water contents from day to day suggests that the balance between the two processes is not critical. The relative contributions to the onset of precipitation by them is presumably quite different on different days, and may well be influenced by the larger-scale patterns in the atmosphere, not necessarily revealed by the tephigrams. More remarkable, and almost certainly *not* explicable in terms of synoptic patterns, is the even greater variation in first-echo heights on a given day (Fig. 18). On 21 July, for instance, one echo (I*b*, Fig. 12) broke out between 13 and 23 kft; another, 32 mi away over similar terrain and only 90 min later (echo II*b*) appeared between 7 and 12 kft. The temperature intervals of these echoes were thus $-8°$ to $-32°$C and $+4°$ to $-6°$C, with correspondingly different liquid-water contents. Variabilities such as these seem to call for meso-scale factors not now understood.

Observations on individual days and especially Fig. 18 suggest a relation between hail and echo height. To examine this statistically, maximum echo-top heights above ground were determined for 47 known hailstorms, and 36 storms without hail; these heights are maximum, with no regard to time of hail-fall. Histograms of these heights are shown in Fig. 19 (a), and the modal value is 10 kft higher for the hailstorms than for

the others. The 83 storms selected occurred on 12 days between 16 July and 20 August, and are probably representative for most of the season, except for the non-hailing low-level storms (tops less than 15 kft) which were more numerous than in our sample. The probability of hail (the ratio of hail-producing storms to the total) shows a clearly increasing trend with echo-top height and to a first approximation increases linearly from zero probability at 10 kft to 100 per cent at 38 kft. Alternatively, one can represent this probability as a function of pressure at echo top, as is done in Fig. 19 (b) and in the equation

$$P = -0{\cdot}25\,(p - p_0)$$

where P is the probability in per cent, and p (in millibars) is the pressure at the echo top. The constant $p_0 = 550$ mb is the minimum value of p for zero probability. Considering the *temperature* at the echo top, the 36 non-hail cases were distributed relatively uniformly across the range of temperatures observed (Fig. 19 (c)). Hail incidence showed a peak in the range -45°C to -49°C; 30 of the 47 cases fell at temperatures below -40°C. At higher temperatures the hail cases were relatively uniformly distributed. These distributions are reflected in Fig. 19 (d), the probability of hail as a function of echo-top temperature; the probability shows no clear-cut trend until the temperature falls below -30°C, but shows a decisive increase at lower temperatures. The relation of hail to echo height, and the absence of a simple relation to echo-top temperature, may well be the most immediately useful findings in this study.

Hail occurrence during the hailstorms was related to the heights of the echo in successive 10 min intervals. On the assumption of a delay of about 20 min between developments near the top of the radar echo and the arrival of hail at the ground, a fairly tight correlation of hail to height of echo top emerged (Fig. 20). But when individual days were examined separately, the two storms of 21 July showed a better correlation of hail arrival to the preceding confluence of echoes than to the preceding echo height. Since it was noted that echo mergers were generally associated with echo upsurges, a relation between hail and echo upsurge seemed to be suggested. However, a statistical examination of our whole sample, comprising nearly 30 hr of storm, has failed to bear this out. Upsurges at modest echo heights do not associate with hail. Using the same data we have also tried to relate temperature at echo top to instantaneous hail falls, but have not found correlations more significant than that based on storms as a whole and shown in Fig. 19.

Our findings about echo heights agree in a general way with the observations in New England of Donaldson (1958), who reported that the probability of hail from a given storm increases with height. It would seem however, that the heights of New England hailstorms are on the whole greater than those in Alberta, and range from 32 kft to 50 kft.

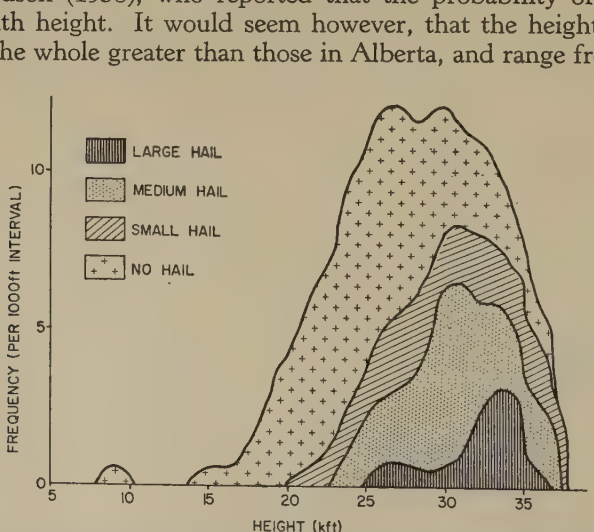


Figure 20. Distribution with height of echo tops of 13 different hail storms during 30 hours (entire area of graph). Inner curves give distribution of hail with height of echo. (An average time-lag of 20 min between echo height and hail arrival was allowed for.) Ordinate scale indicates number of 10-min periods sampled.

A detailed comparison is not possible since Donaldson's deductions (his Fig. 2) are based on point observations of hail and heavy rain, rather than on an areal survey, as was provided by the hail-reporting network in Alberta.

4. Conclusions

A large variety of hailstorms were examined, and the patterns of their radar echoes were related to the hail reports. We encountered high and low storms; some were scattered in time and place, others lasted for 3 hr and longer; on 16 and 23 July the bigger storms covered long and narrow tracks, whereas on 21 July and 20 August the storms tended to be bigger, so that length and width of the tracks were comparable. But in all cases, sustained hail came from storms whose echo heights were steady at heights close to 30 kft above surface. When hail fell while echo tops were lower or less steady, it was not continuous. Short bursts of hail also tended to result when separate echoes merged. On the supposition that echo upsurges reflected the strength of updrafts in the storm a correlation of echo upsurge and hail was sought but could not be found statistically. A more direct investigation of the notion that strong updraughts are required for hail formation (with an updraught speed of 8 m sec^{-1} – the fall speed of rain – possibly critical) is not readily possible, since no good means of measuring local updraughts are available*.

While updraughts may be the important physical parameter, a statistical relation between echo height and hail incidence was discovered. On the basis of an examination of 83 storms occurring in July and August, it was found that the probability of hail falling from a given storm was an increasing function of storm height. An alternative formulation of this relation was given in terms of pressure in § 3. This should be tried for forecasting hail on the basis of a forecast of cumulus tops. Such a forecasting technique may be more immediately promising than an adaptation of the method based on areas on the tephigram which is used successfully in the American Midwest. Though hail incidence relates strongly to height – and possibly to other characteristics of the radar echo, such as horizontal extent – it is not suggested that storm height in itself is the cause of hail; it may merely be a readily measured concomitant of hail, with the primary cause the updraught.

* A radar device for measuring this important quantity could be designed, however. The radar echoes from a storm received at two small neighbouring antennas would be correlated in terms of the fall speed of the echoing particles.

Acknowledgments

The writers wish to acknowledge many stimulating suggestions from all other members of the Stormy Weather Group, and especially from Professor J. S. Marshall, who had a large share in the planning of the project and the organization of the data.

References

Beckwith, W. Boynton. 1956 'Hail observations in the Denver area,' *United Airlines Met. Circ.*, No. 40, Denver, Colorado. (Abridged version : *Bull. Amer. Met. Soc.*, **38**, pp. 20-30, 1957).

Donaldson, R. J. 1958 'Analysis of severe convective storms observed by radar,' *J. Met.*, **15**, pp. 44-50.

East, T. W. R. 1957 'An inherent precipitation mechanism in cumulus clouds,' *Quart. J. R. Met. Soc.*, **83**, pp. 61-76.

Fawbush, E. J. and Miller, R. C. 1953 'A method for forecasting hail size at the earth's surface,' *Bull. Amer. Met. Soc.*, **34**, pp. 235-244.

Foster, D. S. and Bates, F. C. 1956 'A hail size forecasting technique,' *Ibid.*, **37**, pp. 135-141.

Harrison, H. T. 1956 'The display of weather echoes on the 5·5 cm airborne radar,' *United Airlines Met. Circ.* No. 39, Denver, Colorado. (Abridged version : *Aeronaut. Eng. Rev.*, **15**, pp. 102-109, 1956.)

Hitschfeld, W. (Ed.) 1958 'A mechanical-optical system for the fully automatic synthesis of constant-altitude radar maps,' *Tech. Rep. MWT-3A*, being a revision of the earlier *Tech. Rep. MWT-3* by Langleben and Gaherty.

Langleben, M. P. 1956 'The plan pattern of snow echoes at the generating level,' *J. Met.*, **13**, pp. 554-560.

551.524.7 : 551.571.7 : 551.576.11

Observations from aircraft of temperatures and humidities near stratocumulus clouds

By D. G. JAMES
Meteorological Office, London

(Manuscript received 30 June 1958, in revised form 12 November 1958)

Summary

Observations from aircraft of the Meteorological Research Flight flying near stratocumulus clouds showed steep inversions and hydrolapses directly above the tops of the clouds. Turbulence was encountered below and in the clouds and up to 300 ft above the cloud top. Above this level no turbulence was encountered. Analysis of the heat and water-vapour budgets of the cloud and air below suggests that the cloud and the profiles of temperature and humidity can only be maintained if there is large-scale subsidence. The effects of subsidence, outgoing radiation and turbulent transfer are all comparable.

1. Introduction

Winter anticyclones near the British Isles are frequently accompanied by extensive sheets of stratocumulus cloud. Upper-air soundings in these situations show that the tops of such clouds are always limited by steep temperature inversions and sudden decreases in dew-point.

An analysis (James 1955) carried out by the Forecast Research Division at Dunstable of the surface observations and upper-air data on occasions of extensive stratocumulus sheets showed that nearly all cases occurred in the winter months, October to March. Furthermore, the probability that the cloud would persist or dissipate was shown to be related to : (a) The maximum depression of dewpoint below temperature at any pressure level up to 50 mb above cloud top as indicated by the 1500 GMT radiosonde; (b) the average hydrolapse through a 50 mb layer below the cloud base and (c) the cloud thickness.

Clearly the evaluation of the above criteria depends largely on the accuracy with which it is possible to estimate cloud base and top from surface observations and upper-air soundings. Discrepancies in the statistical results could in most cases be directly attributed to ambiguities arising from such estimates, different analysts sometimes placing the cloud sheet at widely differing heights. Furthermore, the lags of the temperature and humidity elements on the radiosonde gave rise to uncertainties as to the magnitude and steepness of the temperature inversion and hydrolapse.

It was thought therefore that a true appraisal of the problem could be obtained only by accurate measurement of some of the features associated with extensive sheets of anticyclonic stratocumulus cloud.

2. Flights by M.R.F. aircraft

Flights carried out from Farnborough (51° 16′N, 00° 46′W) were such that after an ascent to 10,000 ft the aircraft made level runs of 3 to 5 min at 250-ft intervals from 1,000 ft below cloud base to cloud top, then at 100-ft intervals up to 500 ft above cloud top, thereafter at 250-ft intervals to 2,000 ft above cloud top. Height, airspeed, temperature and dew- or frost-point were observed every 250 ft on the ascent and every 30 sec on the level runs. Observations on these level runs were supplemented by accelerometer records. Photographs were taken frequently of the base and top of the cloud sheet.

Eight flights were made in November 1955 but on only four of these flights were full sets of data available. Figs 1 (a) to 8 (b) show the vertical distribution of temperature and dew-point as obtained from the aircraft observations, together with details of cloud base and top.

The second half of November 1955 was particularly favourable for the formation of extensive sheets of stratocumulus cloud over the southern half of the British Isles. An anticyclone centred over Scotland during the early part of the month moved slowly westwards, later returning and moving gradually south-east into the continent. Generally, during this period, south-eastern England was in an anticyclonic, weak north-easterly flow, which was shown by radiosonde soundings to exhibit a well-marked dry-type inversion. Stratocumulus frequently formed under this inversion and sometimes it dissipated during the night. Table 1 gives the dates of the flights together with cloud bases and tops and probable times of dissipation of the cloud.

TABLE 1. FLIGHT AND CLOUD DETAILS

Date	Duration of flight	Cloud Base (ft)	Cloud Top (ft)	Probable time of dissipation
14 Nov. 1955	1400-1630	5,250	5,500	2300/14
15 Nov. 1955	1500-1620	No cloud		—
16 Nov. 1955	1430-1700	3,150	4,000	2200/16
17 Nov. 1955	1520-1620	5,000	6,100	2200/17
18 Nov. 1955	1410-1525	4,200	5,000	Not before 0600/19
21 Nov. 1955	1430-1530	3,000	3,500 (well broken)	Possibly 1800/21
22 Nov. 1955	1340-1630	6,750	7,750	Not before 0600/23
28 Nov. 1955	1100-1330	4,250	4,750	Not before 0600/29

3. TEMPERATURE PROFILES

The temperature measurements were made by a standard Meteorological Office flat-plate thermometer, the lag of which is about 8 sec; a level run of 3 to 5 min thus allows ample time for the thermometer to settle down. On a level run the height of the aircraft was noted to the nearest 10 ft and varied as much as $\pm$ 50 ft so that, particularly in the inversion layer, the temperature fluctuations which were observed would be considerably greater than those which actually occurred at any particular level.

In Figs. 1 (a), 3 (a), 7 (a) and 8 (a) one of the vertical profiles is obtained by plotting the means of the corrected temperatures and dew-points recorded on level runs.

It can be seen that the temperature inversions immediately above cloud top are considerably greater than those shown by the corresponding radiosonde ascents in Figs. 1 (b), 3 (b), 7 (b) and 8 (b).

Figs. 1 (a) to 8 (a) show that: (a) When level runs are made at 100-ft intervals above cloud top the measured temperature gradients in the inversion directly above cloud top are significantly greater than when measurements are made every 250 ft on the ascent; (b) the gradient of temperature with height above the cloud and within 300 ft of cloud top is of the order of 3°C per 100 ft; (c) in some cases the inversion extends as far as 1,000 ft above cloud top, though generally there is a reversal of sign of the lapse rate within 500 ft, and (d) below cloud, the measured temperatures suggest that the lapse rate is approximately dry adiabatic.

4. DEW-POINT PROFILES

The frost-points were measured by a Dobson-Brewer manual hygrometer. In the range in which observations of frost- and dew-points were taken, i.e., $-$ 20°C to $+$ 10°C, this instrument is accurate to about 0·5°C. As the hygrometer is manually operated, the frost-points obtained are not, in general, synchronous with the temperature observations. Also, in a region where the humidity is variable the frost-points are frequently missed by excessive cooling of the observation surface.

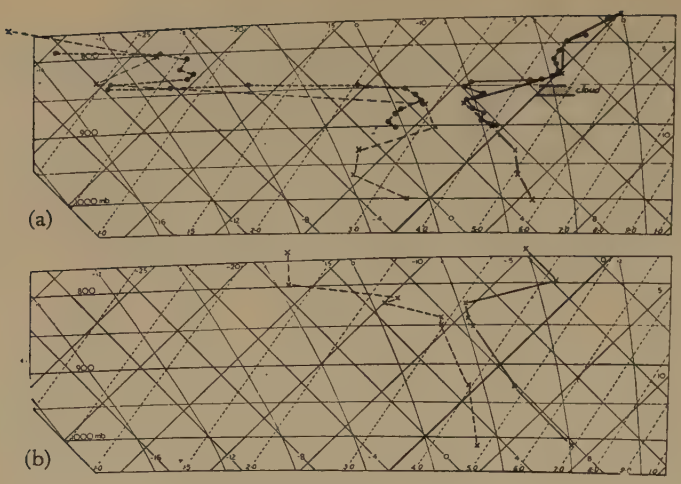


Figure 1. (a) Aircraft measurements, 1400-1630, 14 Nov. 1955; (b) Crawley Radiosonde ascent, 1400 GMT, 14 Nov. 1955.

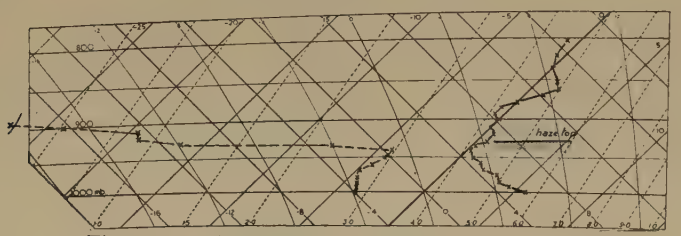


Figure 2. Aircraft ascent, 1500-1620, 15 Nov. 1955.

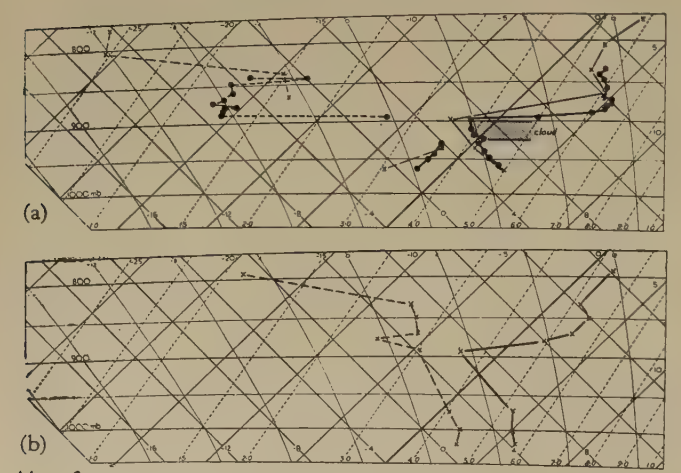


Figure 3. (a) Aircraft measurements, 1430-1700, 16 Nov. 1955; (b) Crawley Radiosonde ascent, 1400 GMT, 16 Nov. 1955.

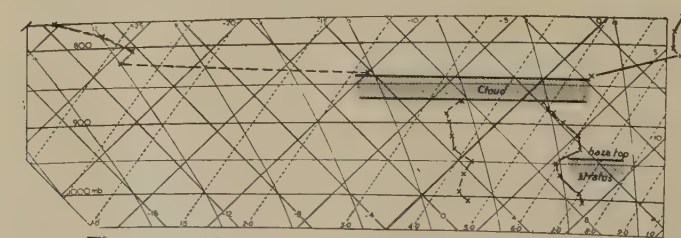
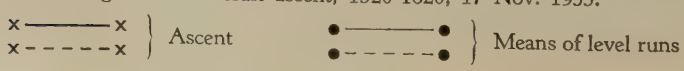


Figure 4. Aircraft ascent, 1520-1620, 17 Nov. 1955.



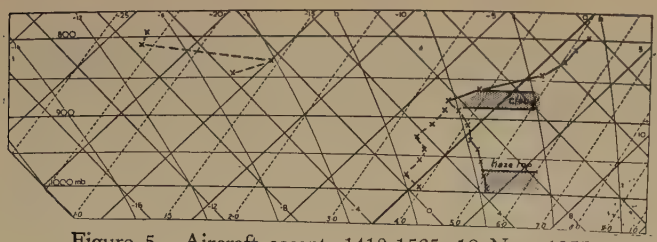


Figure 5. Aircraft ascent, 1410-1525, 18 Nov. 1955.

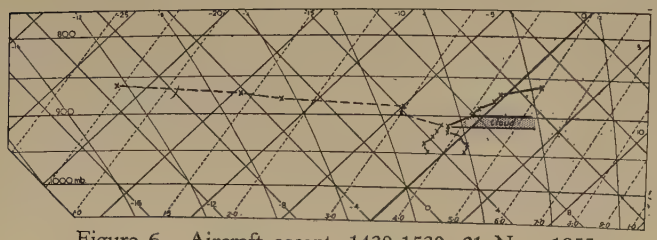


Figure 6. Aircraft ascent, 1430-1530, 21 Nov. 1955.

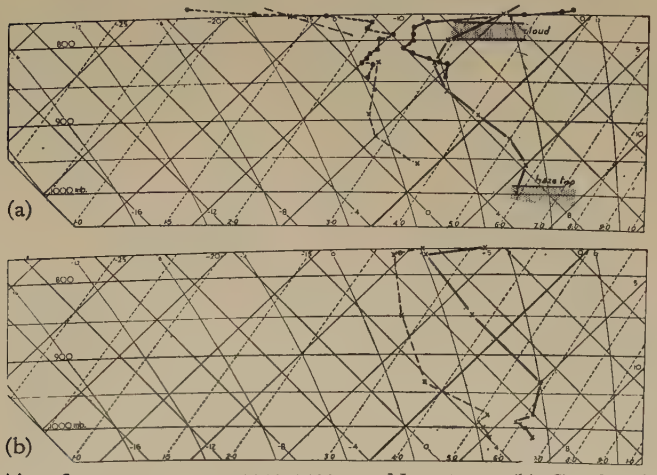


Figure 7. (a) Aircraft measurements, 1340-1630, 22 Nov. 1955; (b) Crawley Radiosonde ascent 1400 GMT, 22 Nov. 1955.

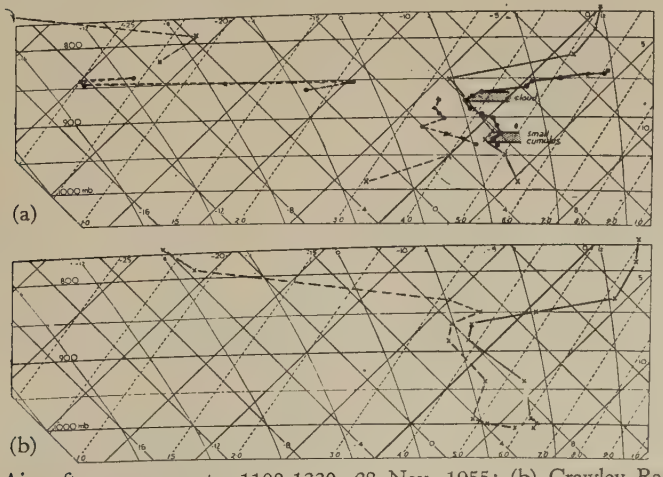


Figure 8. (a) Aircraft measurements, 1100-1330, 28 Nov. 1955; (b) Crawley Radiosonde ascent, 1400 GMT, 28 Nov. 1955.

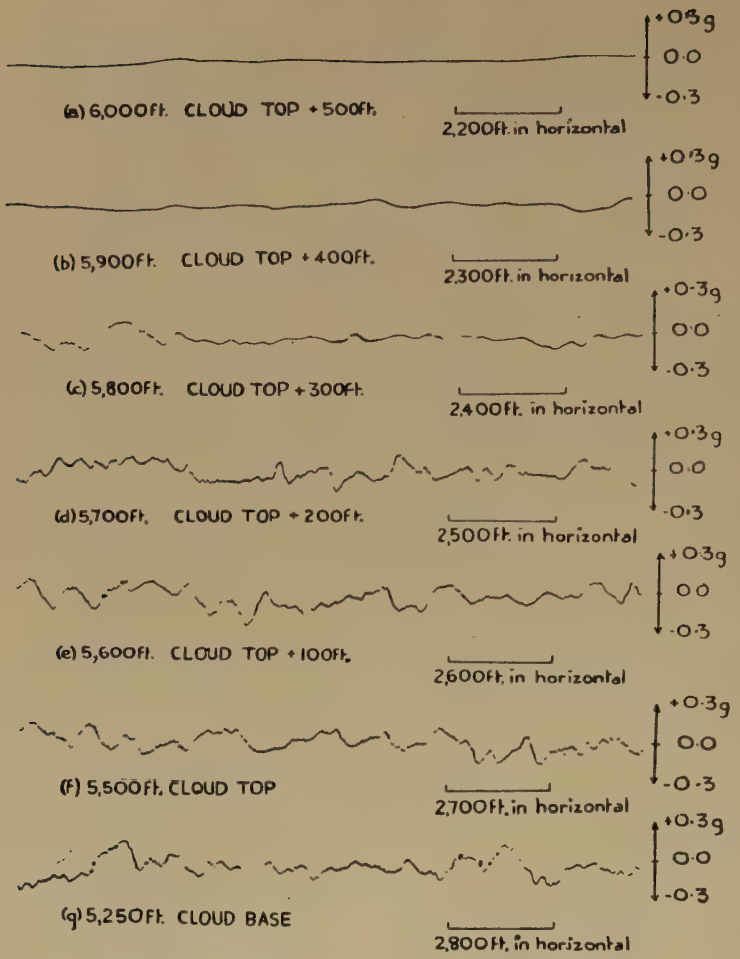


Figure 9. Accelerometer records for 14 Nov. 1955.

Figs. 1 (a), 3 (a), 7 (a), 8 (a) show the vertical profiles of dew-points for the four complete flights. In all cases there is a sudden decrease of humidity just above cloud top, the gradient of humidity mixing ratio with pressure being of the order of 1 g kg^{-1} mb^{-1}. On some occasions the cloud was so thin and tenuous that frost-points were measured in cloud. Nevertheless, complete saturation was not measured.

5. Accelerometer records

Fig. 9 shows sections of the accelerometer traces obtained near stratocumulus cloud on 14 Nov. 1955. The records are similar to those obtained by the aircraft on the other flights.

The traces show that turbulence is experienced below and in the cloud and also up to 200 ft above cloud top. Above this level the number of bumps recorded decreases markedly until at 500 ft above cloud top the aircraft is flying in very smooth air.

6. Discussion

(*a*) *General considerations*

Before considering the heat and water-vapour budgets of the cloud it is necessary to note certain features of the cloud and its environment which are indicated by flight measurements.

Firstly, the cloud sheets had been in existence some 10 hours prior to the times of the flights, and persisted for some hours afterwards; secondly, the profile of temperature and humidity above cloud top are of the same order of magnitude from flight to flight; finally, the turbulence encountered in and below cloud ceases abruptly some 300 ft above cloud top.

The observed turbulence will transfer heat downwards into the cloud top and water vapour upwards into the inversion layer, so tending to change the character of the temperature inversion and hydrolapse. However, since the latter are fairly constant from flight to flight, and in any case indicated no systematic change during any particular flight – albeit of limited duration – then the air in the inversion layer must be continually replaced by subsidence. This downwards transfer of dry, warm air from above the inversion, operates against the turbulence at cloud top to give an approximately steady state at, and immediately above, cloud top.

Before evaluating the rate of subsidence required for this process it is necessary to postulate a rate of mixing at cloud top. This can be achieved by assuming a value for K, the coefficient of eddy diffusion. Taylor (1915) obtains a value of 3×10^3 cm^2 sec^{-1} for K for stable conditions at lower levels over the sea, and this value will be used in the first instance. The effect of a different value of K will be evident at any stage and indeed it will be necessary to consider other values.

In the discussion which follows, the figures given are for the flight of 14 Nov. 1955 (Fig. 1 (a)), but similar calculations have been made for the three other flights and all are presented in Table 2.

(*b*) *The heat budget of the cloud and air below*

Radiation measurements above a sheet of stratocumulus or stratus cloud (Neiburger 1949; Fritz 1950; Murgatroyd 1956) suggest that the albedo of such a sheet varies considerably with cloud thickness. Values between 0·3 and 0·8 have been measured for cloud thickness from ' very thin ' to 4,500 ft. For the cloud investigated on 14 Nov. 1955, which was 500 ft thick, a value of 0·5 will be used.

Measurements at Kew show that on a clear day in November, about 180 cal cm^{-2} are available at the surface from solar radiation, or an average of about 20 cal cm^{-2} hr^{-1} between sunrise and sunset. This figure is supported by data by Charney (1945) and also Houghton (1954) for solar radiation received at the surface at altitude 50°N during November. It will therefore be assumed that about 25 cal cm^{-2} hr^{-1} were available at cloud top i.e. 5,000 ft, on 14 Nov. 1955 from solar radiation. An albedo of 0·5 allows some 13 cal cm^{-2} hr^{-1} to pass into the cloud top.

The loss of heat at the cloud top owing to long-wave radiation, may be obtained by use of the Elsasser radiation chart, and was some 10 cal cm^{-2} hr^{-1} for the cloud considered.

Thus during the day there is a next flux of heat of 3 cal cm^{-2} hr^{-1} downwards through the cloud top, whilst at night there is an outward flux of 10 cal cm^{-2} hr^{-1}.

Measurements at Kew show that, during November, about 1 cal cm^{-2} hr^{-1} is conducted into the earth during the day, the heat being restored at the same rate at night.

Some heat is used in the evaporation of water at the earth's surface. Data available in the Meteorological Office show that the evaporation from tanks of water at Kew was at the rate of about 7×10^{-2} g cm^{-2} day^{-1} on the dates of the flights. It is known that the average rate of evaporation during the day is 3 or 4 times that at night so that these rates are $6{\cdot}0 \times 10^{-3}$ g cm^{-2} hr^{-1} and $1{\cdot}5 \times 10^{-3}$ g cm^{-2} hr^{-1} on the dates of the flights by day and night respectively. The quantities of heat required for these rates of evaporation are approximately 4 cal cm^{-2} hr^{-1} and 1 cal cm^{-2} hr^{-1}.

TABLE 2. ESTIMATES AND OBSERVATIONS ON OCCASIONS OF COMPLETE DATA

Date and time of flight (GMT)	M.s.l. pressure (mb)	Cloud top (mb)	Cloud thickness (mb)	Assumed K $cm^2\ sec^{-1}$	Day 8 night 16 (hours)	Change temp. cloud and air below (°C per day and night)	Change humidity mixing ratio cloud and air below (g kg^{-1} day/night)	Subsidence to keep cloud top at same level (mb hr^{-1})	Expected behaviour of cloud thickness (mb day/night)	Observed behaviour of cloud
14 Nov. 1955				3×10^3	Day	− 0·2	+ 0·2	1	+ 15	Cloud patchy with breaks by day
1400-1630	1025	850	10		Night	− 3·3	− 0·1		+ 40	
Fig. 1 (a)				10^4	Day	+ 0·4	− 0·1	3	− 10	Observed area of cloud probably dispersed 2200 GMT
					Night	− 2·2	− 0·5		− 5	
16 Nov. 1955				3×10^3	Day	0·0	+ 0·2	1	+ 10	
1430-1700	1040	900	25		Night	− 3·7	− 0·1		+ 40	
Fig. 3 (a)				10^4	Day	+ 1·0	− 0·1	3	− 20	Dispersed 2200 GMT
					Night	− 1·7	− 0·7		− 20	
22 Nov. 1955				3×10^3	Day	0·0	+ 0·1	1	+ 5	
1340-1630	1035	780	22		Night	− 2·0	0·0		+ 25	Persisted at least overnight
Fig. 7 (a)				10^4	Day	+ 0·7	0·0	3	− 5	
					Night	− 0·6	− 0·2		0	
28 Nov. 1955				3×10^3	Day	− 0·1	+ 0·2	1	+ 10	
1100-1330	1020	860	12		Night	− 3·4	− 0·1		40	Persisted at least overnight
Fig. 8 (a)				10^4	Day	+ 0·6	− 0·1	2 Heat	− 10	
					Night	− 2·0	− 0·5	3 W. vapour	0	

There is a flux of heat into the cloud caused by turbulent mixing at cloud top. For 14 Nov. 1955, assuming a saturated adiabatic lapse rate in cloud and also a linear variation of temperature with pressure above the cloud top at 851 mb to the top of the inversion at 838 mb, then the average $\frac{\Delta\theta}{\Delta p}$ (where θ is potential temperature) through the stable layer is 0·44°C mb^{-1}.

The flux of heat downwards is thus :

$$F_H = \rho\, c_p\, K \frac{T}{\theta} \frac{\partial\theta}{\partial z} \simeq 1 \text{ cal cm}^{-2} \text{ hr}^{-1}$$

for $K = 3 \times 10^3$ cm^2 sec^{-1}.

The above data show that during the day the net loss of heat by the cloud and air below would be 1 cal cm^{-2} hr^{-1} (see Fig. 10 (a)). Assuming that this loss is uniformly distributed from cloud top to ground – all flights show a dry adiabatic lapse rate of some depth below cloud base – the cloud and air below cool by about 0·2°C during the assumed 8 hours of daylight.

At night there would be a net loss of 9 cal cm^{-2} hr^{-1} giving a cooling of the cloud and air below of 3·3°C during the assumed 16 hours of darkness. This is about twice as great as is actually observed from radio soundings through the same air mass, and must therefore be considered further.

The main factor responsible for this excessive rate of cooling is the long-wave radiation loss from cloud top computed from the Elsasser radiation chart. However, measurements from aircraft of the flux of long-wave radiation near stratocumulus cloud (Brewer and Houghton 1956) confirm the Elsasser value to within $\pm$ 1 cal cm^{-2} hr^{-1}. To obtain a rate of cooling comparable with that observed the cloud and air below should lose heat at about 5 cal cm^{-2} hr^{-1} instead of 9 cal cm^{-2} hr^{-1} calculated above. This could be obtained by increasing the value of K in the inversion layer to 10^4 cm^{-2} sec^{-1} or perhaps a little more. This value of K would, through more rapid transport of heat downwards into cloud top of 4 cal cm^{-2} hr^{-1} restrict the cooling of the cloud and the air below it to about 2·2°C during the night, which is much closer to the actual cooling observed. There is thus some indication that we need to consider values of K considerably greater than 3×10^3 and possibly greater than 10^4 cm^2 sec^{-1}. For a value of K equal to 10^4 the cloud and air below warms by 0·4°C during the day and cools by 2·2°C at night.

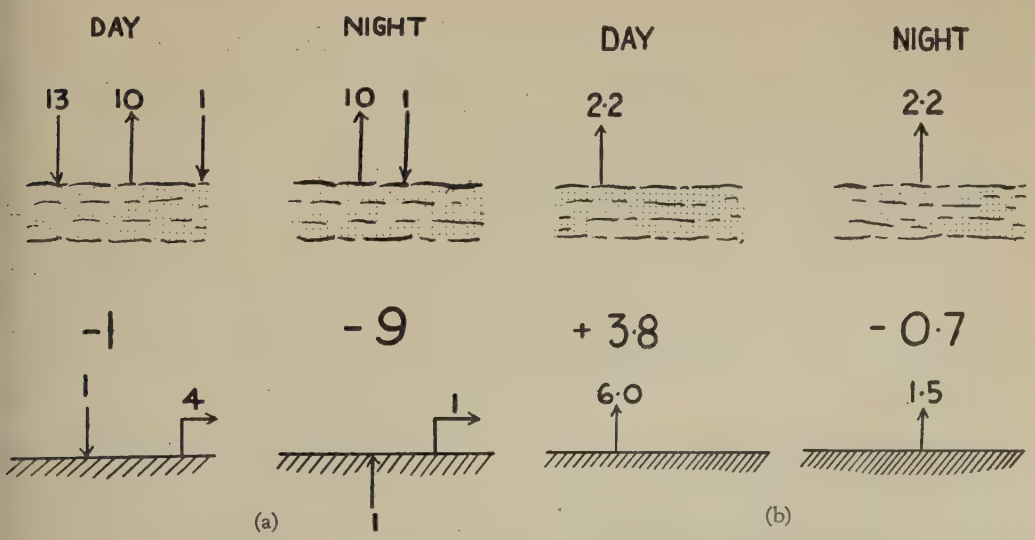


Figure 10. (a) Heat budget of the cloud on 14 Nov. 1955 (cal cm^{-2} hr^{-1}); (b) Water-vapour budget of cloud on 14 Nov. 1955 (10^{-3} gm cm^{-2} hr^{-1}).

(c) *The water-vapour budgets of the cloud and air below*

The water vapour available at the earth's surface has already been given as 6·0 and $1{\cdot}5 \times 10^{-3}$ g cm^{-2} hr^{-1} by day and by night respectively.

The turbulence immediately above cloud results in an upward flux of water vapour by eddy diffusion. The measurement of water content in the cloud is not reliable, but it is reasonable to assume that the humidity mixing ratio in the cloud (including liquid water), is determined by the observed temperature at cloud base and a saturated adiabatic lapse rate through the cloud. Then for 14 Nov. 1955, the average hydrolapse $\frac{\Delta x}{\Delta p}$ through the inversion layer immediately above cloud is $0{\cdot}20 \times 10^{-3}$ g g^{-1} mb^{-1}, so that the upward flux of water vapour is :

$$F = \rho K \frac{\partial x}{\partial z} \simeq 2{\cdot}2 \times 10^{-3} \text{ g cm}^{-2} \text{ hr}^{-1}$$

for $K = 3 \times 10^3$ cm^2 sec^{-1}

The above data mean that during the day there would be a net gain of water vapour of $3{\cdot}8 \times 10^{-3}$ g cm^{-2} hr^{-1} by the cloud and air below (Fig. 10 (b)); by night a loss of $0{\cdot}7 \times 10^{-3}$ g cm^{-2} hr^{-1}. The changes in humidity mixing ratio would thus be + 0·2 g kg^{-1} during the daylight hours and − 0·06 g kg^{-1} by night, K being 3×10^3 cm^2 sec^{-1}. For $K = 10^4$ cm^2 sec^{-1} these changes would be − 0·05 g kg^{-1} and − 0·5 g kg^{-1}, respectively.

(d) *The balance of the inversion layer*

The accelerometer records, Fig. 9, show considerable turbulence in and below the layer from cloud top up to some 10 mb above cloud top, but none whatsoever in the dry air above this level. It must therefore be assumed that the heat gained by the cloud and air below by turbulent diffusion is taken entirely from this layer : also the water vapour lost by the cloud is accumulated in this layer. Now for $K = 3 \times 10^3$ cm^2 sec^{-1}, the heat gained and water vapour lost by the cloud are about 1 cal cm^{-2} hr^{-1} and $2{\cdot}2 \times 10^{-3}$ g cm^{-2} hr^{-1} respectively, so that the inversion layer would cool at about 0·6°C hr^{-1} and its humidity mixing ratio would increase by 0·22 g Kg hr^{-1}. Furthermore, these changes would be immediately effective just above the cloud top where the turbulence is a maximum, and the cloud would grow upwards into the inversion. For $K = 10^4$ cm^2 sec^{-1}, these changes would be 1·3°C hr^{-1} and 0·73 g Kg^{-1} hr^{-1}, respectively. However, no systematic variation in the height of the cloud top was observed on any flight – albeit of limited duration – and so the upward growth of the cloud must be prevented by subsidence, by which the air in the inversion layer is continually being replaced by warm dry air from above.

If air is subsiding through a given layer, then the heat added to that layer is proportional to the rate of subsidence and to the difference between the potential temperature at the limits of the layer. For the data of 14 Nov. 1955 the change in potential temperature through the inversion layer was 6°C, so that for $K = 3 \times 10^3$ cm^2 sec^{-1}, a subsidence rate of rather less than 1 mb hr^{-1} is sufficient to counteract the cooling produced by turbulence.

Similarly, if the transport of water vapour downwards by subsidence is equal to that upwards by turbulence in the inversion layer then a subsidence rate of rather less than 1 mb hr^{-1} will maintain the hydrolapse above cloud top. Thus for $K = 3 \times 10^3$ cm^2 sec^{-1}, a subsidence rate of 1 mb hr^{-1} is sufficient to counteract the turbulent mixing at cloud top and to maintain the cloud top at the same level. For $K = 10^4$ cm^2 sec^{-1}, a subsidence rate of about 3 mb hr^{-1} is required.

(e) *The dissipation of the cloud*

The final colums of Table 2 represent the expected and observed behaviour of the cloud sheets. From these figures it can be seen that the operative value of K in the inversion layer must be at least 10^4 cm^2 sec^{-1} by night if the expected behaviour is to approach the observed behaviour. With a K of 3×10^3 cm^2 sec^{-1} unreasonable rates of subsidence would be required to disperse the cloud in the first two cases by 6 hours after sunset. Conversely, since the cloud in all cases persisted during the day it may be concluded that the value of K at this time is less than 10^4 cm^2 sec^{-1}. This is not an unreasonable hypothesis. It is known that the cooling caused by long-wave radiation from cloud is immediately effective in the uppermost 50 ft or so of the cloud. In the absence of solar radiation, the absorption of which by the cloud would partly offset this loss, the top of the cloud cools very rapidly and produces vigorous convective turbulence inside the cloud. The presence of a dry adiabatic lapse rate below cloud base ensures that this turbulence is effective down to near ground level, but at cloud top the rate of cooling is so great that the convection inside the cloud may well cause overshooting of the cloudy air into the drier air above, thereby increasing the rate of mixing and the value of K. This larger K partly offsets the excessive cooling of the cloud and air below, and also increases the upward flux of water vapour. With a value of $K = 3 \times 10^3$ cm^2 sec^{-1} by day, a rate of subsidence of about 1 mb hr^{-1} is necessary to maintain the level of cloud top. With the increase in K at night this rate of subsidence cannot counteract the upward growing tendency of the cloud, and the cloud top rises, causing an increase in height of the inversion base.

This variation during the night has been noticed by Neiburger (1949) during an investigation of the formation and dissipation of stratus cloud on the west coast of the United States. On several occasions when stratus cloud was present upper-air soundings were made every 3 hours over a period of 48 hours or more. One of the conclusions reached was that the height of the inversion base showed a marked diurnal variation, being a maximum at early morning and a minimum in the evening.

This agrees well with the above assumptions, for if the subsidence rate during the the day and night is rather more than 1 mb hr^{-1} then clearly the base of the inversion will lower during the day without necessarily causing the cloud to dissipate. At night, the base of the inversion will rise by the upward growth of the cloud, until, at sunrise, the long-wave cooling begins to be offset by solar radiation.

7. Conclusions

(a) The temperature inversion and lapse rate of water-vapour content above a sheet of anticyclonic stratocumulus cloud are considerably greater than those shown by radiosonde soundings through the same air mass. Temperature increases of up to 3°C (per 100 ft) and lapse rates of water vapour content of 1·5 g kg^{-1} (per 100 ft) have been measured. For all cases of continuous cloud sheets the change from saturated cloudy air to the warmer drier air occurred almost entirely within the first 300 ft above cloud top.

(b) The accelerometer records obtained from level runs near the cloud show considerable turbulence in and below cloud and also in the first 200-300 ft above cloud top. No turbulence was recorded at 500 ft and greater distances above cloud top.

(c) Consideration of the observed temperature and humidity profiles together with the effects of radiation, turbulent diffusion of heat and water vapour, and subsidence show that the approximately steady state in a persistent sheet of stratocumulus is maintained by subsidence operating against turbulence.

(d) The effective value of K, the coefficient of eddy diffusion immediately above cloud top is of the order of 10^4 cm^2 sec^{-1} and probably greater by night than by day. With $K = 10^4$ cm^2 sec^{-1}, subsidence of 3 mb hr^{-1} would keep cloud top at the same level in the cases considered. Subsidence rates greater than 3 mb hr^{-1} would result in cloud top falling and conversely.

(e) A factor which materially affects the heat and water balance of the cloud is the increase in K which occurs in the absence of solar radiation. This may lead to the dissipation of a cloud at night which had been persistent during the day.

8. Acknowledgments

I am indebted to Mr. R. H. Clements for much helpful advice during the preparation of this work. I also wish to thank the Director-General of the Meteorological Office for permission to publish this paper.

References

Brewer, A. W. and Houghton, J. T.	1956	*Proc. Roy. Soc.* A, **236**, p. 175.
Charney, J.	1945	*Radiation* (Handbook of Met.), New York. McGraw Hill, p. 284.
Fritz, S.	1950	*Bull. Amer. Met. Soc.*, **31**, p. 251.
Houghton, H. G.	1954	*J. Met.*, **11**, p. 1.
James, D. G.	1955	Air Ministry, Met. Res. Ctee., M.R.P. No. 913.
Murgatroyd, R. J.	1956	*Ibid.*, No. 995.
Neiburger, M.	1949	*J. Met.*, **6**, p. 28.
Taylor, G. I.	1915	*Phil. Trans. Roy. Soc.*, A215, p. 1.

551.558.21

Theory of mountain waves of large amplitude

By R. S. SCORER and H. KLIEFORTH*
Imperial College, London

(Manuscript received 10 June 1958)

SUMMARY

Rotors are defined as regions containing flow in the opposite direction to the main stream and are shown to exist when the wave amplitude is large enough. Equations are derived for computing rotor flow in an infinitely deep atmosphere.

The problem of choosing an appropriate second boundary condition in steady flow over a ridge is complicated when the amplitude is finite because the shape of obstacle obtained depends on the airstream characteristics as well as upon the mathematical form given to it. Probably under some actual conditions the oncoming airstream is modified by the presence of the mountain.

Because of the great variety of airstreams and mountains no general case can be illustrated. The particular cases chosen illustrate rotor flow. These particular cases do, however, each represent a set of cases in which all the non-dimensional numbers involved have the same value.

Known cases of rotors in the Sierra Nevada and at Cross Fell are cited to illustrate the theory.

1. INTRODUCTION

Long (1953) showed that the equations for two-dimensional flow in a stably stratified liquid were linear in certain cases, even when the amplitude was large. It was possible to explain rotors in this way. The purpose of the present paper is to apply these methods to the atmosphere. The main difficulty arises from the fact that in Long's (1953, 1955) experiments, the stratified liquid was effectively contained between two rigid surfaces and was set in lee-wave motion although it was composed of only one layer. It is now well known that in order to have lee waves in the atmosphere the parameter l^2 must vary with height in a suitable way. If l varies continuously with z the equations for waves are not linear when the amplitude is large. If l is discontinuous, and we have a multilayer model in each layer of which l is constant, the conditions at the interfaces are not linear and cannot in general be satisfied. This difficulty has been overcome by choosing special cases in which the linear equations are valid even for large amplitude. This means that some, at least, of the characteristics of large-amplitude waves can be more or less correctly studied. We have not investigated the fact that the solution of the non-linear equations is not periodic in x, except possibly in very special cases, and the methods are only valid for two-dimensional flow.

The notation follows that used by Scorer (1955).

It was proved (Scorer 1955, Eq. 5.7) that provided the displacement of a particle from its original level was such that

$$\epsilon \ll 1, \quad . \quad . \quad . \quad . \quad . \quad (1.1)$$

where

$$\epsilon = g\zeta/c^2 \quad . \quad . \quad . \quad . \quad (1.2)$$

and c is the velocity of sound, then the vertical displacement ζ in two-dimensional motion in the x-z plane satisfies

$$\nabla^2 \zeta + \left(\frac{\beta_0}{2} + \frac{g}{c^2} - \frac{U'}{U}\right)(\nabla\zeta)^2 + \left(\frac{2U'}{U} - \beta_0 - \frac{g}{c^2}\right)\frac{\partial\zeta}{\partial z} + \frac{g\beta_0}{U^2}\zeta = 0 \quad . \quad (1.3)$$

β_0 and U, the static stability and wind speed in the stream when undisturbed, are constant along a streamline and are therefore functions of x and z which depend upon ζ: the equation is therefore non-linear in ζ in all terms except the first unless U and β_0 are independent of z, in which case only the second term is non-linear. If we can choose a special

* Now at Geophysics Research Directorate, U.S. Air Force, Cambridge Research Center, Bedford, Mass.

case in which the equation is linear without being restricted to waves in which the streamlines are not far from horizontal we can discover some of the properties of waves in which the velocity is at right angles to, or in the opposite direction to, that in the undisturbed stream. In perturbation theory it is assumed that u and w are small compared with U; in arriving at (Eq. 1.3) it is only necessary to set a limit to ζ.

2. The condition for the existence of a rotor

A rotor exists when the flow in part of the fluid is in the opposite direction to the main stream. If there is no mixing across the streamlines, and ψ is the stream function constant along the streamlines such that in the undisturbed stream $\partial\psi/\partial z > 0$, then in a rotor $\partial\psi/\partial z < 0$. This means that in a rotor

$$\frac{\partial\zeta}{\partial z} \gtrless 1 \quad \text{according as} \quad \zeta \gtrless 0 \qquad . \qquad . \qquad . \qquad (2.1)$$

Where $\partial\zeta/\partial z = \pm 1$ (with $\zeta \gtrless 0$ respectively) the flow is vertical. If this happens at the ground there is a point of separation of the flow from the surface. Since the ζ-profile may assume a variety of shapes there is a corresponding variety of rotor types. These are illustrated in Fig. 1 for the case of a wave crest or trough, in which the crests lie in vertical planes. On the boundary of the region in which the flow is upside down $|\partial\zeta/\partial z| = 1$. The boundary of the closed circulation which is the rotor, is the streamline through the extreme points on which the flow is vertical. The rotor occurs only in the wave crest if $\partial\zeta/\partial z > 0$ and in the trough if $\partial\zeta/\partial z < 0$. There may be a double-yolked rotor at a nodal surface.

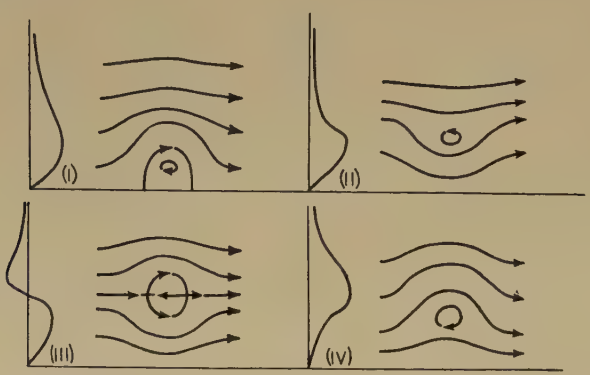


Figure 1. On the left of each diagram is shown the ζ-profile, and on the right the corresponding wave form when the amplitude is large enough for rotors to occur. The rotor occurs at the level of maximum $\partial\zeta/\partial z$. The rotors are closed circulations with the main flow passing round them.

If a rotor exists in a stable airstream a statically unstable lapse-rate is produced where the direction of flow is reversed. Although flow strictly along the steady-state streamlines would be possible, any disturbance will lead to some sort of mixing of the air within the rotor until it all has the same potential temperature. In such a case if condensation were to occur in the top half of a rotor, very great instability would be produced because the surroundings of the cloud would have an adiabatic lapse rate : except near a hot surface where vertical motion is inhibited by the surface, greater buoyancies are probably never produced naturally anywhere else in the atmosphere. Fragmentary cloud within a rotor can be expected to have exceptionally great vertical motion. The case of a glider being broken apart in such cloud is referred to later.

If the wave crests or troughs are tilted a rotor may be open with streamlines in the shape shown in Fig. 2. Static instability is produced in air which flows through the pattern so that steady motion could not persist. Exactly what happens can only be guessed

Figure 2. If the wave amplitude *and phase* change suitably with height an *open rotor* may be produced. In all rotors the air becomes statically unstable and therefore liable to mixing by buoyant convection. If this happens in an open rotor the mixed air must be carried along with the main flow so that the configuration of flow shown must be unsteady. Theoretically, open rotors may or may not have closed circulations in the bends of the flow. Such a circulation could not be realised in practice in steady flow.

at present, but if the wave pattern were re-established each time the static instability produced some mixing across the streamlines and the mixed fluid were carried away downstream some sort of pulsating motion would seem likely. If it were visible through the formation of cloud or through the crumpling of a sharp haze top illuminated by a low sun, it would look somewhat like a breaking wave, and the existence of this kind of rotor may sometimes be responsible for billows in stably stratified air. This matter is probably best pursued experimentally and will not be discussed further here.

3. Rotors in lee waves

The number of conditions which can be varied is so great that a general formal solution would not be useful even if it could be obtained. Particular cases can be chosen to illustrate the dynamical properties of disturbed stable airstreams. The most interesting airstreams from the practical point of view are those which are set into lee-wave motion downstream of any disturbance. The simplest case is that in which the lee waves are periodic in the horizontal direction with wavelength $2\pi/k$. In general, solutions of (Eq. 1.3) with the boundary conditions for lee waves cannot be periodic in x unless the equation is linear in ζ, and this means that if a term is not negligible its coefficient (which is constant along a streamline) must be independent of z. In order to discuss a model airstream we shall therefore choose one composed of layers in each of which

$$\left.\begin{aligned} & g\beta_0/U^2 = l^2 = \text{constant}, \\ & U'/U \;\; = S = \text{constant} \\ \text{and either} \quad & (i) \quad U'/U \gg g/c^2, \beta \\ \text{or} \quad & (ii) \quad \text{the term in } \partial\zeta/\partial z \text{ is negligible.} \end{aligned}\right\} \quad . \quad . \quad (3.1)$$

This last alternative condition is such that if (*i*) holds, the coefficient of $\partial\zeta/\partial z$ can be regarded as independent of z because its major component is. If (*i*) does not hold, then in order to make Eq. (1.3) linear the term in $\partial\zeta/\partial z$ must be negligible in comparison with the other terms, and this is true if $U' = 0$ because the terms in β_0 and g/c^2 are negligible (and have generally been ignored in previous work).

We have also to arrange that the term in $(\nabla\zeta)^2$ shall be negligible : this means that either $\partial\zeta/\partial z$ must be sufficiently small for the term in $\partial\zeta/\partial z$ to dominate that in $(\nabla\zeta)^2$, or the coefficient of $(\nabla\zeta)^2$ must be negligible. To achieve this we arrange that $\partial\zeta/\partial z$ is comparable with unity only in the regions where $U' = 0$. The terms in β_0 and g/c^2 are again negligible.

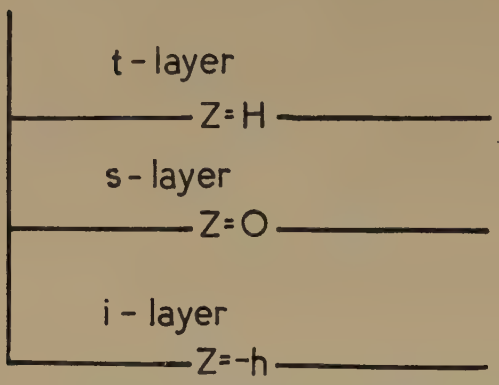


Figure 3. Notation for 3-layer model.

Under these conditions the equation becomes

$$\frac{\partial^2 \zeta}{\partial z^2} + 2S\frac{\partial \zeta}{\partial z} + (l^2 - k^2)\,\zeta = 0 \quad . \quad . \quad . \quad (3.2)$$

If we consider an airstream composed of three layers denoted by suffixes i (inferior), s (superior), and t (top), which are such that $l_i = \text{const} > k$, S_i and S_t are of the order of g/c^2 or less;

$$l_s = \text{const}, \quad S_s = S = \text{const}, \quad l_t = \text{const} < k, \quad . \quad . \quad (3.3)$$

then ζ_i satisfies
$$\left.\begin{aligned} &\zeta_i'' + \nu^2 \zeta_i = 0 \\ &\nu = +(l_i^2 - k^2)^{\frac{1}{2}} \end{aligned}\right\} \quad . \quad . \quad . \quad . \quad (3.4)$$

ζ_s satisfies
$$\zeta_s'' + 2S\zeta_s' + (l_s^2 - k^2)\,\zeta_s = 0 \quad . \quad . \quad . \quad (3.5)$$

and ζ_t satisfies
$$\left.\begin{aligned} &\zeta_t'' - \mu^2 \zeta_t = 0 \\ &\mu = +(k^2 - l_t^2)^{\frac{1}{2}} \end{aligned}\right\} \quad . \quad . \quad . \quad . \quad (3.6)$$

Solutions of these equations are

$$\zeta_i = C(\cos \nu z + \cot \nu h \sin \nu z) \quad . \quad . \quad . \quad (3.7)$$

$$\zeta_s = C\,e^{-Sz}(n \sin qz + \cos qz) \quad . \quad . \quad . \quad (3.8)$$

$$\zeta_t = C\,e^{(\mu - S)H}(n \sin qH + \cos qH)\,e^{-\mu z} \quad . \quad . \quad (3.9)$$

where
$$q = +(l_s^2 - k^2 - S^2)^{\frac{1}{2}} \quad . \quad . \quad . \quad . \quad (3.10)$$

and n is to be determined.

In these it has been arranged that ζ is a continuous function of z, thus

at
$$\left.\begin{aligned} &z = 0 \text{ (lower interface) } \zeta_i = \zeta_s = C \\ &z = -h \text{ (the ground) } \zeta_i = 0 \\ &z = H \text{ (upper interface) } \zeta_s = \zeta_t = C\,e^{-SH}(n \sin qH + \cos qH) \end{aligned}\right\} \quad . \quad (3.11)$$

When perturbation theory is used, the conditions that the pressure and vertical displacement are continuous across any interface between layers in which the solution is known are applied at the level of the interface in the undisturbed condition, and are therefore only strictly satisfied at points where the interface is at its original level (and ζ is zero). If there is no discontinuity of density then since $gz + \frac{1}{2}|v|^2 + \varpi/\tau$ is constant along a streamline (Scorer 1955) continuity of ϖ implies continuity of $|v|$ (speed along the streamline) if there is no discontinuity of $|v|$ at one point on the interface. This means

that ζ' is continuous at the interface (because the spacing of the streamlines is determined by ζ'). The problem therefore is to arrange that ζ and ζ' shall be continuous at the interface, which may be displaced by perhaps as much as 1 km.

If the conditions are satisfied at the level of the undisturbed position of the interface they will be well satisfied at the actual interface provided that the amplitude of the disturbance is not large and that ζ'' is not large. However, in our examples, the amplitude at the lower interface will be large, while to be linear, the conditions must be imposed at a fixed z. In order to ensure that the conditions are not seriously violated at the displaced position we apply the best possible conditions at the undisturbed position, namely that ζ and ζ' are continuous, and that ζ'' is also continuous. Higher derivatives of ζ cannot be made continuous without the conditions becoming too stringent to be useful. Consequently since

$$\left.\begin{aligned} \zeta_i' &= \nu C\,(-\sin \nu z + \cot \nu h \cos \nu z) \\ \zeta_s' &= C\,e^{-Sz}\{(qn - S)\cos qz - (Sn + q)\sin qz\} \\ \zeta_t' &= -C\,\mu\,e^{(\mu - S)H}\,(n \sin qH + \cos qH)\,e^{-\mu z} \end{aligned}\right\} \quad . \quad . \quad (3.12)$$

at $z = 0$, where $\zeta_i' = \zeta_s'$, we find that

$$\nu \cot \nu h = qn - S \quad . \quad . \quad . \quad . \quad . \quad . \quad (3.13)$$

and at $z = H$, where $\zeta_s' = \zeta_t'$, we find that

$$\cot qH = \frac{n(S - \mu) + q}{nq - (S - \mu)} \quad . \quad . \quad . \quad . \quad . \quad (3.14)$$

Also
$$\left.\begin{aligned} \zeta_i'' &= \nu^2 C\,(\cos \nu z + \cot \nu h \sin \nu z) \\ \zeta_s'' &= C\,e^{-Sz}\{(S^2 - 2n\,Sq - q^2)\cos qz + (nS^2 + 2Sq - nq^2)\sin qz\} \end{aligned}\right\} \quad (3.15)$$

so that since $\zeta_i'' = \zeta_s''$ at $z = 0$, we find that

$$-\nu^2 = S^2 - 2n\,Sq - q^2 \quad . \quad . \quad . \quad . \quad . \quad (3.16)$$

Eq. (3.13) determines n as a function of k if the characteristics of the airstream are given, but so also do Eqs. (3.14) and (3.16). This means that there exist restrictions on our choice of airstreams. We can exercise choice in determining the values of l_i, l_s, l_t, S, h, and H, so that there is no inherent difficulty in choosing an airstream for which Eqs. (3.13), (3.14) and (3.16) are true, but owing to the involved way in which k occurs in the equations it is not practicable to eliminate it formally and discover the whole class of airstreams which satisfy these requirements. Of the whole class, only part would satisfy the further requirements that the amplitude at the upper interface should be small and that $(U'/U)\,(\nabla\zeta)^2$ should be negligible in the middle layer when the amplitude is large enough for there to be a rotor in some part of the flow. We have therefore chosen the wavelength to be 2π km approximately, for convenience, and sought for model airstreams which satisfy the other conditions. The flow calculated for these examples is discussed later.

4. Choice of solution

There has been some discussion (e.g., Scorer 1954; Corby and Sawyer 1958; Scorer 1958; Palm 1958) on how to choose a unique solution of the wave equation, which in this case is Eq. (1.3) or its modified form Eq. (3.2). For the purpose of drawing the pattern of streamlines in the diagrams of this paper we have used the solution suggested by Scorer (1958). Because it is the nature of the large-amplitude lee waves which we wish to illustrate and the precise choice makes little difference to this part of the diagram we shall only discuss here the effect of finite amplitude.

When the amplitude is large, the shape of the mountain obtained by these methods is not determined by the choice of ζ at a particular value of z alone, but also by the airstream. For instance if we put $\zeta = \zeta_1(x)$ at $z = 0$, and the airstream above has a value of

l independent of height, then the solution obtained with this boundary condition and Scorer's choice of solution will be, with the usual approximations for a medium-width mountain

$$\zeta = \zeta_1(x) \cos lz + \text{lee waves} \quad . \quad . \quad . \quad . \quad (4.1)$$

If there are no lee waves the mountain shape will be given by $\zeta = z$, i.e.

$$\zeta_1(x) = z \sec lz \quad . \quad . \quad . \quad . \quad . \quad (4.2)$$

which has not the same shape as the small-amplitude solution which is

$$\zeta_1(x) = z \quad . \quad . \quad . \quad . \quad . \quad . \quad (4.3)$$

When there are lee waves which are present only for $x > 0$ they have the effect of making the mountain asymmetrical even if $\zeta_1(x)$ is an even function of x. This has already been noted by Long (1955). If we wish to draw the flow pattern over a symmetrical mountain $\zeta_1(x)$ must be unsymmetrical. When the lee waves are of large amplitude, as for instance in the Sierra waves in the Owens Valley in California, we do not need to have a function ζ_1 as unsymmetrical as the mountain shape to be represented.

For a uniform airstream without lee waves the general solution with the usual approximations and with small amplitude is

$$\zeta = \zeta_1(x) \cos lz + f(x) \sin lz \quad . \quad . \quad . \quad . \quad (4.4)$$

Scorer suggests that $f(x) = 0$. In that case when the amplitude is large it is easy to choose $\zeta_1(x)$ to give any required mountain shape for we only have to make (4.2) the equation of the mountain section. But if we choose $f(x) = \bar{\zeta}_1(x)$, where $\bar{\zeta}_1$ is a function derived from ζ_1 by the application of a ' boundary ' condition (so called, but usually at infinity) then in order to find what function ζ_1 is we have to solve

$$z(x) = \zeta_1(x) \cos lz(x) + \bar{\zeta}_1(x) \sin lz(x) \quad . \quad . \quad (4.5)$$

where $z = z(x)$ is the given mountain shape. In that case the functional relationship between ζ_1 and $\bar{\zeta}_1$ cannot be independent of x, and the relationship derived from the boundary condition cannot be used as it has been for mountains of infinitesimal amplitude unless the mountain shape happens, by chance, to be one which gives a relationship between ζ_1 and $\bar{\zeta}_1$, which is independent of x and agrees with the boundary condition.

If the boundary is ignored then there are two functions, ζ_1 and $\bar{\zeta}_1$, one of which can be chosen arbitrarily, and the other of which is then determined by Eq. (4.5).

Scorer's argument was that $f(x)$ should be zero because its coefficient $\sin lz$ is zero at the level of the mountain ($z = 0$, in this case for infinitesimal amplitudes) and it is not produced by the mountain unless another boundary condition in finite space or time is applied. An extension of this idea to mountains of large amplitude would require that the solution be

$$\zeta(x, z) = \zeta_1(x) \cos l(z - z^*(x)) \quad . \quad . \quad . \quad (4.6)$$

where the mountain shape is given by $z = z^*$ and also by $\zeta = z$, and so we simply obtain

$$\zeta_1(x) = \text{mountain height} \quad . \quad . \quad . \quad . \quad (4.7)$$

This would not apply to the lee-wave part of the solution, and we have not used this solution here.

With any of the solutions of Eq. (4.4), which includes all the small perturbation solutions, there are heights at which $\zeta = 0$. This implies that we cannot have this kind of flow over mountains which exceed this height. With solutions of Eq. (4.1) the height cannot exceed π/l. Long (1955) noted some interesting consequences of confronting a stream with an ' impossible,' or nearly impossible, mountain in his model experiments. Before mentioning these it is noteworthy that this problem does not arise if Eq. (4.6) is accepted as the solution for an infinitely deep uniform airstream. The other solutions do not suggest how such an airstream could flow over a mountain of height greater than

π/l at $x = 0$. Eq. (4.6) would, of course, not apply to Long's experiments because the upper rigid boundary at finite height makes it inappropriate, but it suggests that the problem can only be solved by applying the boundary condition at the height of the mountain, and not by trying to obtain the mountain by applying the condition at another place. Long found that the stream was modified either upstream by the passage of a wave upstream from the mountain, reducing the value of l, or there was a hydraulic jump modifying the stream permanently on the lee side; this happens if there are no waves which can travel upstream to modify the oncoming flow because the stream is too rapid. Equally the flow might not occur in the manner supposed : for instance, the lower layers might be blocked and become stationary with the upper layers descending to the surface on the lee side (this was not possible in Long's experiments). Consequently we cannot suppose that any airstream will flow across any mountain – still less that its flow can be represented by the present means even when the equations are linear. Whether any of these things happen in the atmosphere has not been investigated; but the obvious suggestion is that föhn winds will occur if an airstream reaches a mountain, say on the arrival of a cold front, whose height exceeds π/l.

We have seen that a good approximation to lee waves of large amplitude could be obtained by arranging for Eqs. (3.13), (3.14) and (3.16) to be satisfied. It was not practicable to eliminate k from these equations and so it was arranged that they should be satisfied for the value of k corresponding to stationary (lee) waves. We have also seen that the existence or otherwise of rotors is not dependent upon whether the equations are linear, the chief effect of non-linearity being to make strictly periodic waves usually impossible. Over a ridge only one or two wavelengths wide, the effect of the non-linearity would not, therefore, have much effect; and so we have computed the flow over a ridge with lee waves and rotors as if the linear equations, which are good for the lee waves, are good for the flow over the ridge also.

The present theory cannot be applied except to two-dimensional flow for reasons given by Scorer (1955).

5. Three-layer models of lee waves with rotors

In the construction of possible two-dimensional models, the problem is to select values of the airstream characteristics ν, μ, h, and S which will satisfy the following conditions :

(1) The maximum amplitude ζ_{max} should be of the order of 1 km to satisfy Eq. (1.1) when the amplitude, C in Eq. (3.7) etc. is chosen so that $\zeta'_{-h} = 1$, and the condition (2.1) for separation of the flow and incipient rotor development at the ground is just satisfied.

(2) At $z = H$, ζ should not be more than about 0·2 km, for at larger amplitudes the condition $\zeta_s = \zeta_t$ is not sufficiently well satisfied at the disturbed interface, even though it is at $z = H$.

(3) There are to be no nodal surfaces in the middle layer because ζ' has its maxima near to, or at, the zeros of ζ so that even if there were no rotors at the nodal surface the term in $(\nabla\zeta)^2$ might become comparable with that in ζ' in Eq. (1.3) and the form (3.2) would be incorrect.

To simplify the calculations we choose $\nu = 1$ and use the kilometre as the unit of length. A convenient value for the wavelength, L, is 6 km so that $k \simeq 1$. As will be shown in § 6, a particular solution fitting the above requirements can be scaled to give a corresponding solution for any other wavelength.

To find a suitable model airstream, the value of $\cot \nu h$ was varied from $+2$ to -2. The range of S is determined by the condition $q^2 > 0$ in Eq. (3.10). The values of μ and S were increased until the condition $\zeta_H \simeq 0{\cdot}2$ km was satisfied. The results are summarized in Fig. 4. It is seen that as the i-layer increases in depth the s-layer decreases. Characteristics of each of the five cases are briefly noted :

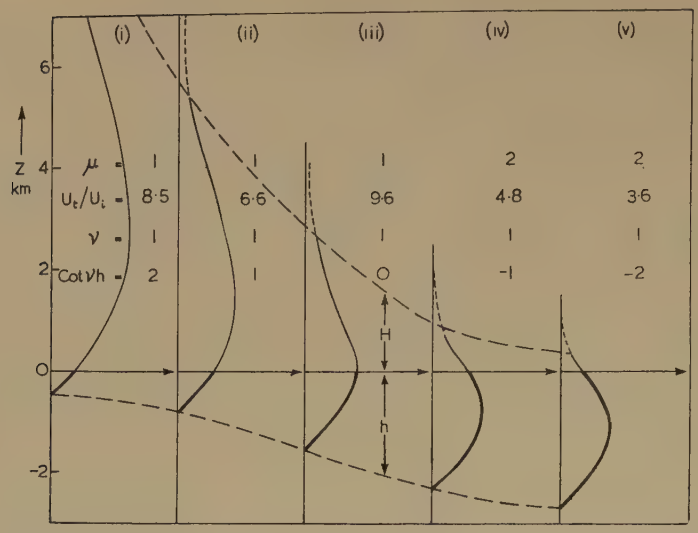


Figure 4. The five model airstreams investigated, satisfying Eqs. (3.13), (3.14) and (3.16). The profile of ζ is shown in each case. A heavy line is used in the i-layer, and a dashed line in the t-layer. In each case $\partial\zeta/\partial z = 1$ at $z = -h$.

(*i*) $\cot \nu h = +2$. h is shallow, less than 0·5 km in depth. $H = 9{\cdot}75$ km so that a shearing layer of great stability extends to what would be approximately the level of the tropopause. In order to satisfy $\zeta'_{-h} = 1$, $\zeta_{\max}$ must exceed 1·5 km

(*ii*) $\cot \nu h = +1$. The s-layer is 5·4 km thick and the ratio U_t/U_i is 6·6. $\zeta_{\max} \simeq 1{\cdot}1$ km and occurs 1·5 km above $z = 0$ (2·3 km above the ground).

(*iii*) $\cot \nu h = 0$. $\zeta_{\max} = 1$ km and is at the interface $z = 0$. $h = 1{\cdot}57$ km and $H = 2{\cdot}62$ km.

(*iv*) $\cot \nu h = -1$. The i-layer of constant wind speed U_i has a depth of 2·4 km. $\zeta_{\max}$ occurs within this lowest layer. While H is equal to only 0·8 km, the wind shear in the s-layer is great and it has been necessary to increase μ to a value of 2 thus making $L \simeq 3$ km.

(*v*) $\cot \nu h = -2$. The i-layer is rather deep ($h = 2{\cdot}7$ km), the s-layer is very shallow ($H = 0{\cdot}3$ km), the wind shear is big ($S = 2{\cdot}0$), and the necessary wavelength ($L \leqslant 2$ km) is very short.

The two most realistic cases, (*ii*) and (*iii*) above, were selected for the study of rotor development.

6. Construction of lee-wave streamlines

In the models chosen the equations for the displacement of a streamline reduce to the form.

$$\zeta = Cf(z) \sin kx$$

or

$$x = \frac{1}{k} \sin^{-1} \frac{\zeta}{Cf(z)} \quad . \quad . \quad . \quad . \quad (6.1)$$

$Cf(z)$ is given in Eqs. (3.7), (3.8) and (3.9). Now $\zeta = z - z_0$, where z_0, the label of a particular streamline, is the level of that streamline in the undisturbed flow. For a given z_0, z is varied and the corresponding values of x are computed.

The streamlines are spaced so as to be inversely proportional to the wind speed. In the lowest and uppermost layers they are therefore equally spaced in the undisturbed state, while in the middle layer the spacing decreases upwards.

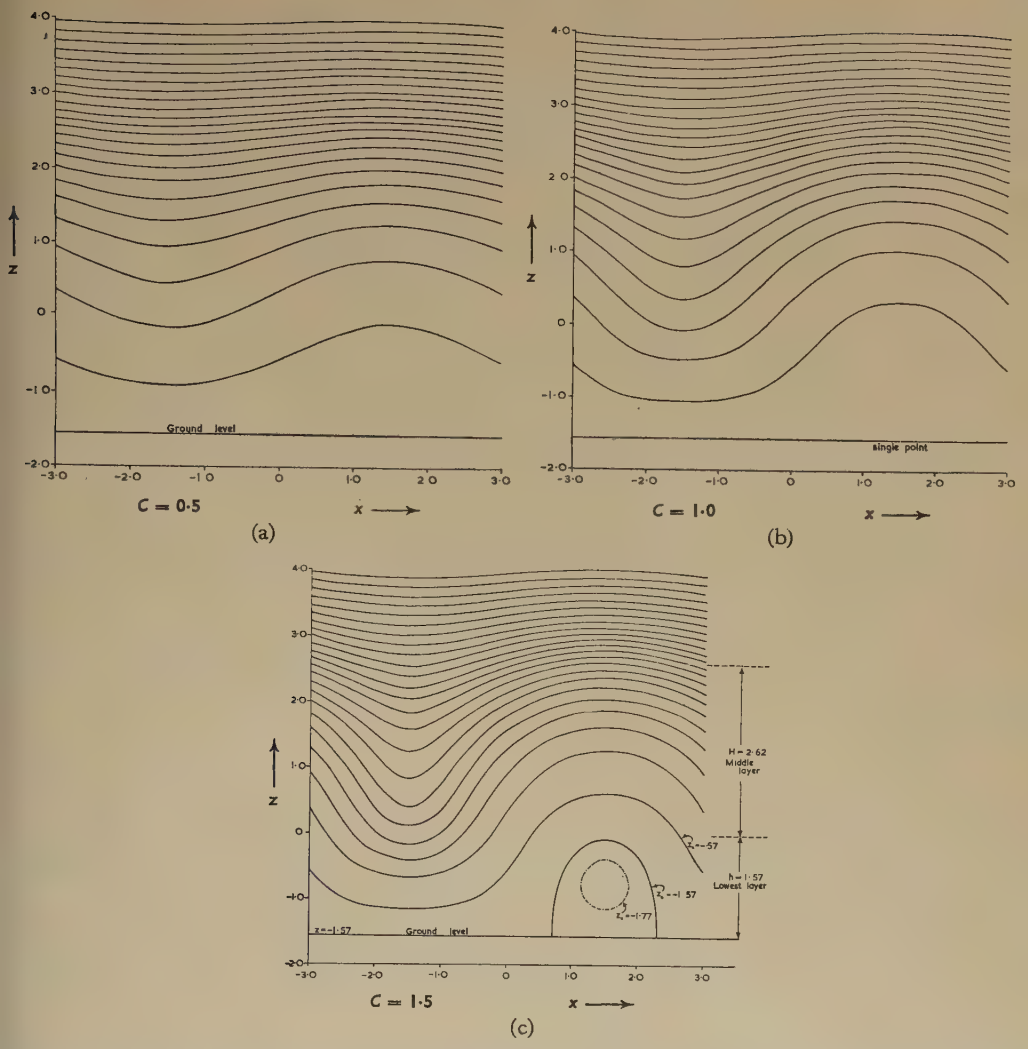


Figure 5 (a), (b), (c). The streamlines in a large-amplitude lee-wave for three values of the amplitude with the same formal solution of the equations. For sufficiently large amplitude reversed flow and a rotor appears at the ground with sinusoidal wave flow at high levels.

For each model the streamlines were constructed for 3 values of C, giving values of $\partial\zeta/\partial z$ equal to 0·5, 1·0 and 1·5 at the ground. Those for case (*iii*) are shown in Fig. 5. When $(\partial\zeta/\partial z)_{\max} = 0{\cdot}5$ there is no rotor. When it is increased to 1·0 there is one point at the ground under the wave crest at which the velocity is zero. With $(\partial\zeta/\partial z)_{\max} = 1{\cdot}5$ a well-developed rotor appears, with reversed flow at the ground.

Each solution represents a family of model airstreams. For example, we obtain the same pattern using the parameters σh, σH, σL, ν/σ, μ/σ, S/σ in the place of L, H, L, ν, μ, S with U_i and U_t kept the same. But to have a rotor, i.e., to make $\zeta'_{-h} \geqslant 1$, we must then have amplitude σC. σ is a numerical multiplier. x and z become σx and σz in the diagrams.

In our three-layer model l^2, $= g\beta/U^2$, is constant for each layer and therefore the stability β is proportional to U^2. To obtain other cases we may replace the velocity by σU and the stability by $\sigma^2\,\beta$; the pattern then remains unaltered. In this case σ may be a function of z provided its variations are not large, and although the shape of the streamlines is unaltered, their spacing may be. Now β, $= \theta^{-1}\,\partial\theta/\partial z$, increases with z in the

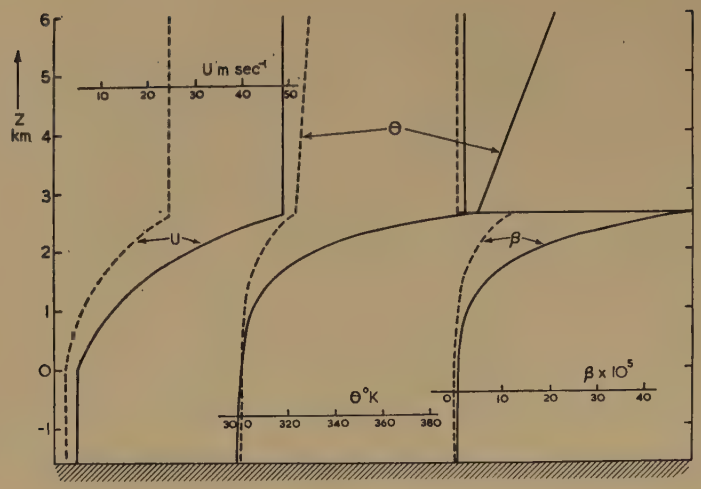


Figure 6. Wind and temperature profiles of possible airstreams whose streamline shape is illustrated in Fig. 5. In one $U_i = 2{\cdot}5$ m sec^{-1} in the other $U_i = 5$ m sec^{-1}. The corresponding profiles of β and θ are shown on the right. β is constant in the i- and t-layers and is discontinuous at $z = H$. The θ-profiles were computed assuming that $\theta = 300°$K at $z = 0$.

middle (s-) layer and the potential temperature sounding in this layer can be plotted from the formula

$$\ln \theta = \frac{U_i^2 l_i^2}{2gS} e^{2Sz} + \text{const.} \quad . \quad . \quad . \quad (6.2)$$

where the constant is determined by the value of θ at $z = 0$. From Eq. (6.2) it can be seen that strong shear and high wind speeds are associated with a deep, very stable layer with its top at the level $z = H$.

In the real atmosphere there is generally no such correlation between vertical wind shear and increasing stability with height. In the best wave situations at least, l^2 usually decreases continuously with z in an airstream with a strong wind speed maximum at high levels even though β may be only slightly greater than zero. However, airstreams with high wind speeds often contain strong frontal inversions and these layers, with their high stability and strong vertical wind shear, are analogous to the s-layer of the model.

In Fig. 6 the relationship between the velocity and temperature profiles is illustrated for one of the cases treated above. In the case of our models we find in general, that when attempting to simulate the observed wind profile of an actual airstream with lee waves, the static stability of the sounding is much less than that of our model.

7. Airflow over the mountain

Having got airstreams that will form lee waves of large amplitude, the next step is to determine the form of the mountain ridge that will excite such waves. Bearing in mind the remarks of § 4, the complete equation for airflow over a ridge given by Scorer (1953) is taken as

$$\zeta = ab \frac{\zeta_{z,0}}{\zeta_{1,0}} \frac{b}{b^2 + x^2} - 2\pi ab \frac{\zeta_{x,k}}{\partial \zeta_{1,k}/\partial k} e^{-kb} \sin kx \quad . \quad . \quad (7.1)$$

where the subscript 1 refers to the ground, a is the height of the mountain, and b is its half-width. The first term on the right describes the flow in the neighbourhood of the mountain; the second term describes the lee waves, and is absent when $x < 0$.

To obtain the largest-amplitude waves for a given mountain height, we choose $b = 1/k = 0{\cdot}955$ km. We have in case (*iii*)

$$\zeta_{z,k} = \cos\left[(l_i^2 - k^2)^{\frac{1}{2}} z\right] \quad . \quad . \quad . \quad . \quad . \quad (7.2)$$

$$\zeta_{1,k} = \cos\left[-(l_i^2 - k^2)^{\frac{1}{2}} h\right] \quad . \quad . \quad . \quad . \quad (7.3)$$

$$\partial\zeta_{1,k}/\partial k = hk\,(l_i^2 - k^2)^{-\frac{1}{2}} \sin\left[(l_i^2 - k^2)^{\frac{1}{2}} h\right] \quad . \quad . \quad (7.4)$$

Now $k = 1{\cdot}047$, $h = -\pi/2$, and $(l_i^2 - k^2)^{\frac{1}{2}} = \nu = 1$. So $\partial\zeta_{1,k}/\partial k = 1{\cdot}643$. Also at $z = 0$, $\zeta_{z,k} = 1$ and $\zeta_{z,k} \Big/ \dfrac{\partial\zeta_{1,k}}{\partial k} = 0{\cdot}609$.

When $\sin kx = 1$ and $z = 0$, the second term on the right in Eq. (7.1) is made equal to 1·0, the critical value of C for rotors to occur. This gives $a = 0{\cdot}744$ km. When $C = 1{\cdot}5$ and b remains the same, $a = 1{\cdot}10$ km (which is ζ at the level of the plain below the mountain crest).

To determine the pattern of flow over the mountain we use Eqs. (3.7), (3.8) and (3.9). One then sets $k = 0$ and the displacements of streamlines in the three layers are :

$$\zeta_i = A\left\{e^{il_i(z+h)} + B\,e^{-il_i(z+h)}\right\} \quad . \quad . \quad . \quad . \quad (7.5)$$

$$\zeta_s = e^{-Sz}\left\{C\,e^{iq_0 z} + D\,e^{-iq_0 z}\right\} \quad . \quad . \quad . \quad . \quad (7.6)$$

$$\zeta_t = E\,e^{il_t z} + F\,e^{-il_t z} \quad . \quad . \quad . \quad . \quad . \quad (7.7)$$

where q_0 is the value of q for $k = 0$.

According to § 4, $B = 1$. From the conditions :

$$\text{at} \quad z = 0, \quad \zeta_i = \zeta_s \quad \text{and} \quad \zeta_i' = \zeta_s'$$

$$\text{at} \quad z = H, \quad \zeta_s = \zeta_t \quad \text{and} \quad \zeta_s' = \zeta_t'$$

the remaining arbitrary constants are then determined in terms of A. As explained at the end of § 4 we assume that these conditions, satisfied at $z = 0$ and $z = H$, are near enough satisfied at the disturbed positions of the interfaces. Then

$$\zeta_i = 2A\cos l_i\,(z + h) \quad . \quad . \quad . \quad . \quad . \quad (7.8)$$

$$\zeta_s = A\,e^{-Sz}\,(c_1\cos q_0\,z - c_2\sin q_0\,z) \quad . \quad . \quad . \quad (7.9)$$

$$\zeta_t = A\,(c_3\cos l_t\,z + c_4\sin l_t\,z) \quad . \quad . \quad . \quad . \quad (7.10)$$

where c_1, c_2, c_3, and c_4 are real numbers dependent on the airstream characteristics. A does not appear in the ratio $\zeta_{z,0}/\zeta_{1,0}$ of Eq. (7.1). Letting $r = \zeta_{z,0}/\zeta_{1,0}$, and noting that $\zeta = z - z_0$, the streamlines are plotted from

$$x = \pm\, b\,(ar/\zeta - 1)^{\frac{1}{2}} \quad . \quad . \quad . \quad . \quad . \quad (7.11)$$

The pattern of the flow over a mountain with the lee waves omitted (the first term of (7.1)) for the case (*iii*) has a downward displacement of the streamlines of the middle layer above the mountain crest and a consequent increase in the wind speed. Above a nodal surface at about 2·4 km there is an upward displacement of the streamlines. If the value of S were decreased, the displacement of the upper streamlines would be increased. In the case illustrated in Fig. 7, the disturbance above the mountain is less than that in the lee waves.

The flow over an antisymmetric mountain profile can be studied by choosing the profile

$$\zeta_1 = -\frac{abx}{b^2 + x^2} = \text{imaginary part of } -ab\int_0^\infty e^{-kb+ikx}\,dk \quad . \quad (7.12)$$

The first term in (7.1) then becomes $-ab\,\dfrac{\zeta_{z,0}}{\zeta_{1,0}}\,\dfrac{x}{b^2 + x^2}$. The airflow over such a profile is also antisymmetric. Thus at some upper level over a mountain range with a steep leeward escarpment there will be a streamline at height π/l that is a mirror image of the ground

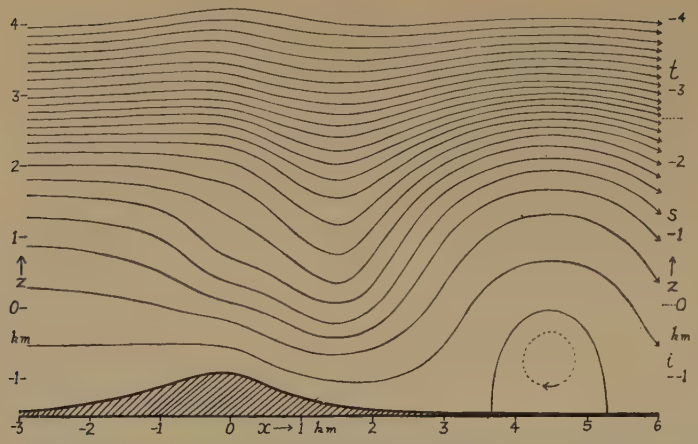


Figure 7. The streamlines for the flow over a two-dimensional ridge of the airstream whose lee waves are shown in Fig. 5. Theoretically there would be an infinite train of waves of which only the first one is shown. However, the rotor would necessarily be turbulent so that the second lee wave would probably be of smaller amplitude. As explained in section 6 this flow pattern, though particular, could occur in a restricted class of airstreams, but qualitatively it is like what might occur in a much wider class. The chief restriction is that the airstream can have *periodic* lee waves of large amplitude.

profile. If the air is sufficiently moist for condensation to occur in the steep upward sweep of the streamline, a cloud will appear at high levels in the lee of the mountain crest, and it may have been many miles upwind of the crest that cloud evaporated in the air approaching the mountain.

The flow over the mountain has been combined with the lee waves of case (*iii*) to give the flow over a ridge with lee waves and is shown in Fig. 7. On account of the lee waves, the mountain itself becomes asymmetrical because of its finite amplitude (as in Long's cases). It must be emphasized that this is a special case in the sense that it is the flow over a particular mountain of a particular airstream in which strictly periodic lee waves are possible; but a set of flow patterns can be deduced from it as explained in section 6. Nevertheless, the object in displaying it is to show the existence and nature of rotor flow, to emphasize in particular that turbulent flow in the rotor with vertically upward velocities at the front of it can occur with smooth wave flow above in an infinitely deep atmosphere. Such flow is illustrated in the colour photograph by Miss Betsy Woodward, published in *Cloud Study* (1957) and in *Weather*, Dec. 1957.

8. Observations of rotor phenomena

In England, Manley (1945) has described in detail the helm wind of Cross Fell on the western slope of the Pennines. This is a strong local wind in the stable easterly flow of a winter anticyclone. One observes over the crest of the range a cap cloud (helm cloud) and downwind of the foot of the slope one or more roll clouds (helm bars). The strong downslope wind is experienced at the ground downwind of the cap while under the roll cloud a reverse flow is often observed. Manley emphasized the commonness of sharp inversions just above the rotor cloud. If the lapse rate were near adiabatic below the inversion, with $l_i > k$, the maximum value of $\partial\zeta/\partial z$ might occur in the free air just below the inversion and not at the ground (see, for instance, cases (4), (6) and (7) of Fig. 2, Scorer 1954). The rotor could then be free from the ground as illustrated in Fig. 1 (iv) of this paper. Reversed flow in the rotor could then occur without reversed flow at the ground.

In the Owens Valley of eastern California rotors frequently develop under the crests of lee waves in airstreams flowing over the Sierra Nevada. Usually these circulations are invisible but occasionally they are nicely shown by dust or smoke. At times a point of separation is marked by alkali dust rising vertically upwards and spiraling over; at other

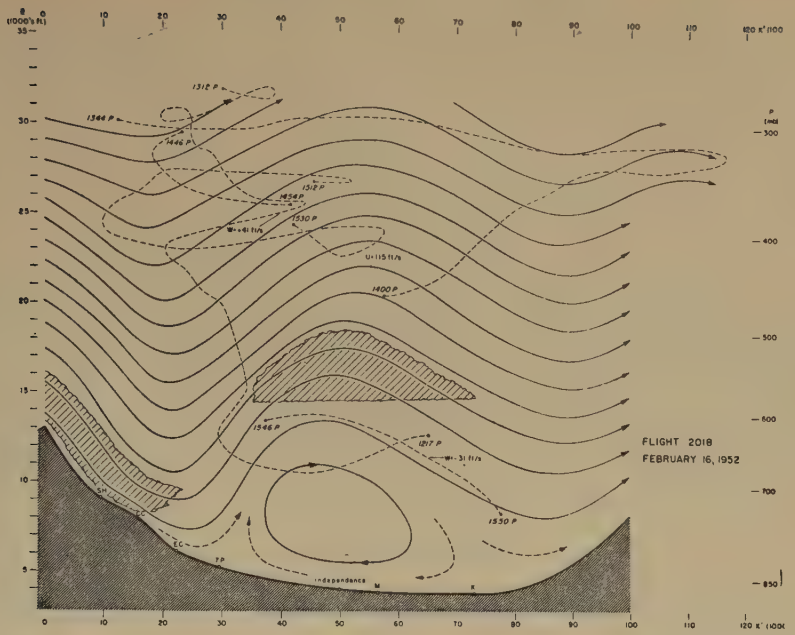


Figure 8. Streamline pattern deduced from observations of rotor flow in the Owens Valley on 16 Feb. 1952, made by the Sierra Wave Project.

times, when rotors are large and the lapse rate near the ground is nearly adiabatic, dust-whirls in the downslope and reversed flows have been seen to converge toward a line below the leading edge of the roll cloud.

An example of a strong lee wave with rotor cloud is shown in Fig. 8. This is a vertical cross-section in which streamlines have been drawn from measurements made on 16 Feb. 1952 by a sailplane of the Mountain Wave Project (Holmboe and Klieforth 1957). The roll, or rotor, cloud, which is well above the level of reverse flow, derived its name from the illusion of rotation caused by the big vertical wind shear between its base and top. The depth of the reversed flow in this case is approximately 1,300 m. The top of the roll cloud is often a few thousand feet higher than the top of the cap cloud over the mountains.

On another occasion (25 April 1955) a unique observation of the rotor circulation was made when the Sierra Wave Project sailplane broke apart in severe turbulence near the upwind edge of the roll cloud and the pilot, Mr. Larry Edgar, descended through this region by parachute. After being carried rapidly down the direction of the main stream eastward across the Valley below the roll cloud, he encountered a layer of calm air at about 2,500 m (1,300 m above the ground) below which he drifted westward in a wind estimated at 25 knots. He finally landed on the *west* side of Owens Valley below the leading edge of the roll cloud.

References

Corby, G. A. and Sawyer, J. S.	1958	*Quart. J. R. Met. Soc.*, **84**, p. 25.
Holmboe, J. and Klieforth, H.	1957	Sierra Wave Project, Final report, March 1957, (Dept. of Met., Univ. of California).
Long, R. R.	1953	*Tellus* **5**, p. 42.
	1955	*Ibid.*, **7**, p. 341.
Ludlam, F. H. and Scorer, R. S.	1957	*Cloud Study : a Pictorial Guide*, John Murray, London.
Manley, G.	1945	*Quart. J. R. Met. Soc.*, **71**, p. 197.
Palm, E.	1958	*Ibid.*, **84**, p. 484.
Scorer, R. S.	1953	*Ibid.*, **79**, p. 70.
	1954	*Ibid.*, **80**, p. 417.
	1955	*Ibid.*, **81**, p. 551.
	1958	*Ibid.*, **84**, p. 182.
Woodward, B.	1957	*Weather*, **21**, No. 12.

551.558.1

The motion in and around isolated thermals*

By BETSY WOODWARD
Imperial College, London†

(Manuscript received 15 August 1958; in revised form 3 Jan. 1959)

SUMMARY

Laboratory measurements show that the motion of an isolated 'thermal' is like that of a vortex ring. The centre is rising at slightly more than twice the rate of the thermal cap and on the edges the buoyant fluid is descending, relative to a stationary observer. As the thermal rises, mixing with outside air takes place, approximately 60 per cent at the front and 40 per cent at the rear, and the thermal traverses a cone of half angle about 15°. Heavy particles, with a terminal velocity of less than 0·4 W_c, where W_c is the rate of rise of the cap, will not escape from the thermal; whilst those with a terminal velocity exceeding somewhat less than 1·6 W_c cannot be retained in it. It is argued that the isolated thermal is the ultimate form of all convective elements in the atmosphere and that this ultimate form is generally reached between 200 and 2,000 metres from the ground.

1. INTRODUCTION

The experimental technique used and the preliminary results obtained in laboratory experiments on studies of the convection of isolated masses of buoyant fluid have been reported by Scorer (1956, 1957, 1958) and Scorer and Ronne (1956).

Salt solution, made visible by a white precipitate, was poured into a thin-walled hemispherical cup placed at the top of a 5 ft × 3 ft × 3 ft tank of water. The cup was quickly inverted by hand, allowing the salt solution to sink down from rest through the water in the tank. Initially the fluid accelerated but after travelling about 1·5 diameters deceleration began.

As the purpose of the present paper is to draw the analogy between the experiment and the rising of warm thermals in the atmosphere the experimental results will be described as though the motion was that of lighter fluid moving upward through a medium rather than that of a heavier fluid descending. The vertical coordinate is inverted for that purpose.

By simple dimensional analysis it is found that in this decelerating phase

$$W = C\,(g\,\bar{B}\,R)^{\frac{1}{2}} \qquad . \qquad . \qquad . \qquad . \qquad (1)$$

$$Z = n\,R \qquad . \qquad . \qquad . \qquad . \qquad . \qquad (2)$$

where W is the vertical velocity of the foremost part of the buoyant (in this case heavy) fluid, $\bar{B}$ the mean buoyancy of the fluid defined by $\dfrac{\rho_0 - \rho_i}{\rho_0}$ where ρ_i is the mean interior density and ρ_0 the exterior density, R is the horizontal radius of the cloud and Z the distance to the foremost part of the cloud from a virtual origin. The constants C and n were determined from experiment to be about 1·2 and 4·0 respectively.

The total excess buoyancy $\bar{B}\,V$, where V is the volume of the buoyant fluid, remains constant when the exterior density ρ_0 is uniform; thus $\bar{B}$ is given at any time by the relation

$$\bar{B}\,V = B_0\,V_0 \qquad . \qquad . \qquad . \qquad . \qquad . \qquad (3)$$

where B_0 is the initial buoyancy and V_0 the initial volume.

Writing $W = dz/dt$ and integrating Eq. (1) we obtain

$$k\,Z^2 = t \qquad . \qquad . \qquad . \qquad . \qquad . \qquad (4)$$

where k is a constant for a given thermal, but which is different for different thermals.

* Presented at the 7th OSTIV Congress, Leszno, Poland, June 1958.

† Now with Arthur D. Little, Inc., Cambridge, Mass., U.S.A.

The way in which it depends upon the total buoyancy of the thermal has been discussed by Scorer (1957, 1958).

The virtual origin, or point source, was obtained by plotting R against Z, extrapolating backwards and taking $Z = 0$ where $R = 0$.

2. The general motion

The general motion of the buoyant fluid, which we shall call a thermal, is rather like that of a vortex ring. Inside and at the rear of the thermal the flow is relatively smooth. Mixing with the outside fluid takes place at the forward surface, or cap. This mixing takes place by the thermal growing protuberances in between which the exterior fluid is captured. Each protuberance looks like a small thermal and has smaller protuberances upon it. The limit to the smallness is presumably set by viscosity.

Figure 1 shows the successive outlines of two thermals. Protuberances move around the edge of the thermal from the cap. As a result the ratio R/Z fluctuates. Even though the thermal may sometimes grow asymmetrically (a) a fairly constant value of k is nevertheless maintained.

Because the thermal has symmetry about a vertical axis, the distribution of motion in and around the thermal is completely specified if it is determined in any axial plane. Experimental measurements of this motion are presented in this paper, together with some deductions.

3. Experimental technique

The thermal was illuminated through its vertical axis by a flat beam of light 2-3 cm thick and the motion of suspended particles, or tracers, within it recorded on 16 mm ciné film. Small polythene beads, made visible by a coat of white paint were used as tracers; they were about 4 mm in diameter and had a small terminal velocity in water.

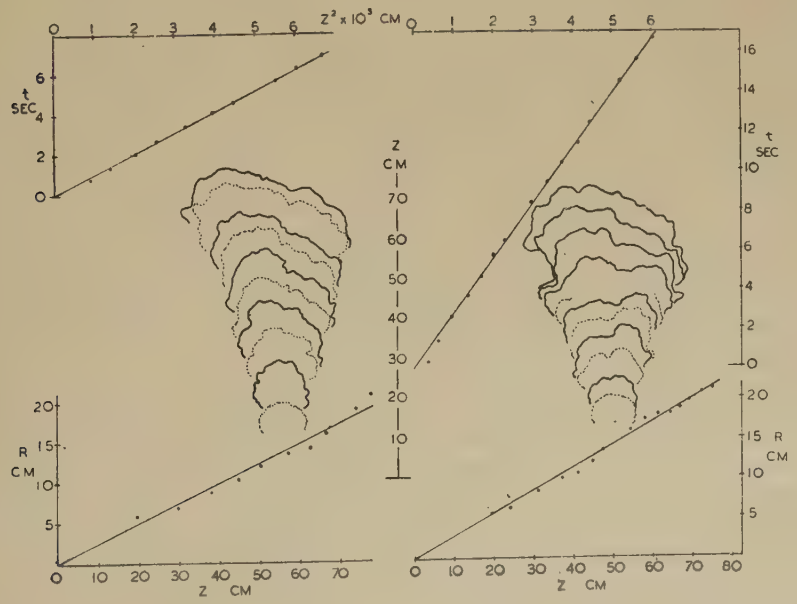


Figure 1. Successive outlines of thermals traced from photographs. Below each is a graph of height, Z, against radius, R, and above each Z^2 is plotted against t. Even though the ratio R/Z fluctuates, owing to protuberances moving around the edge, a fairly constant value of t/Z^2 is maintained.

In the early experiments this velocity was taken into account when computing velocities due to the motion of the thermal. In later experiments the particles had a terminal velocity in water of less than 5 mm sec^{-1} which was negligible in comparison with the velocities produced by the thermal. The latter particles were released from a horizontal thread lying in the beam of light about 40 cm ahead of the cap.

4. Reduction of data

From each frame of the ciné film the following measurements were made : time, the distance of the thermal cap from a fixed reference, the radius of the thermal, and the co-ordinates of the particles.

Since the values of n and k are taken to be constant for each experiment, the motion is *similar* at all stages. This implies that in and around the thermal the velocity at any point expressed as a fraction of the velocity of the cap of the thermal is a function only of the coordinates of the point expressed as a fraction of the height of the cap. That is

$$\frac{w}{W} = f_1\left(\frac{z}{Z}, \frac{r}{Z}\right) \qquad \frac{u}{W} = f_2\left(\frac{z}{Z}, \frac{r}{Z}\right)$$

where w and u are the instantaneous vertical and radial velocities at position z, r and f_1 and f_2 are undetermined functions.

The fluid velocity was measured from the motion of the particles. From successive frames separated by a time interval Δt, w and u were determined from the relations

$$w = \frac{\Delta z}{\Delta t} \qquad u = \frac{\Delta r}{\Delta t}$$

where Δz is the vertical displacement, Δr the radial displacement of the particle during Δt.

The values of w and u thereby obtained were divided by the appropriate value of W $\left(\text{determined from the relation } W = \frac{1}{2kZ}\right)$ and plotted on a diagram using z/Z, r/Z as co-ordinates.

5. Results

The similarity of the motion at different stages in the history of a thermal was again confirmed and lines joining equal vertical and radial velocities were then drawn.

These are shown in Fig. 2 and are the average results of five different experiments. The motion will be described as if the thermal were rising. It can be seen that the vertical velocity in the centre is about 2·2 times the vertical velocity of the cap, and that around the edge of the thermal, fluid is descending at slightly more than half the speed at which the cap is rising. The dashed lines denote radial velocities : outflow in the upper part of the thermal, inflow in the lower. The thermal also has an appreciable effect on the surrounding fluid : the fluid above is ' pushed ' up before it becomes engulfed, there is downward motion at the edge of the thermal, and some of the fluid to the rear flows in and up. As the thermal rises its radius increases, approximately following the cone of half angle 15°.

The greatest difficulty experienced in obtaining the results shown in Fig. 2 was the measurement of n. This is because as the larger protuberances advance around the edge of the thermal the radius fluctuates. In the experiments chosen to compute the velocities shown here, only those which had a fairly constant value of n were selected.

A variation in n may also occur even though the thermal grows along a fixed cone. This is because of the change in the average shape of the thermal. Thus, for instance, if two thermals with values of n equal to 4·8 and 4·2 follow a cone of half angle 15°, the maximum radius of the first occurs at a lower point relative to the cap than in the second case.

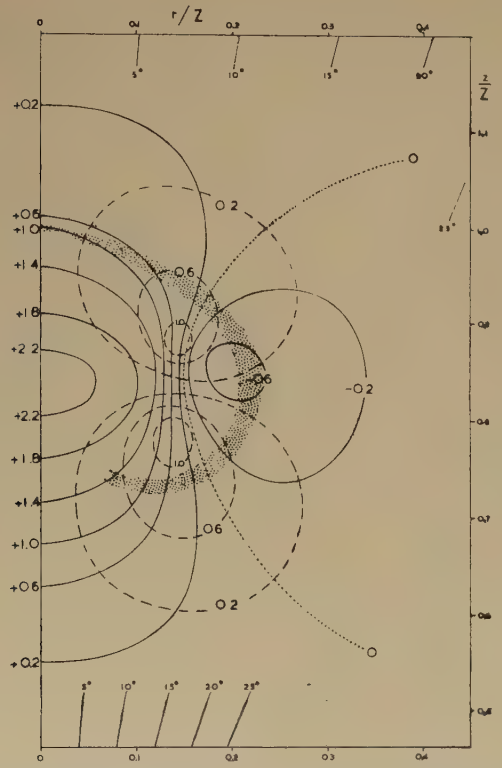


Figure 2. The distribution of velocity in and around an isolated thermal. Only the right-hand side of the thermal is shown. The outline of the buoyant fluid is shaded. The values of the vertical velocities (solid lines) and radial velocities (dashed lines) are expressed as multiples of the vertical velocity of the thermal cap.

The magnitudes of average velocities shown in Fig. 2 can be taken to be correct. Their position, however, may vary about by 15 per cent of the radius. This allows for a variation of n with time in a relatively symmetrical thermal. If the thermal grows asymmetrically a greater error of position will occur.

We have discussed average velocities only. At the leading edge of the thermal there is a region (the thickness at the centre of the thermal being about 15 to 20 per cent of the radius) where mixing occurs. The velocity of individual particles in this region varies considerably from the average values.

Fig. 3 shows successive positions of particles expressed in terms of their relative co-ordinates (z/Z, r/Z) obtained from actual measurements of two experiments.

The region of turbulence near the cap is shown by the irregularity of the paths of the particles.

Fig. 3 may be compared with Fig. 4 which shows the paths of particles of fluid relative to the thermal and computed from the average velocities given in Fig. 2. If the actual displacements of a fluid element in time interval Δt are Δz vertically and Δr radially, then the changes in its relative co-ordinates are

$$\frac{\Delta z - \frac{z}{Z}\Delta Z}{Z} \quad \text{and} \quad \frac{\Delta r - \frac{r}{Z}\Delta Z}{Z}$$

The first term in each quantity expresses the change in relative position owing to the motion of the fluid element. The second term takes into account the changing relative position due to the growth of the thermal. These expressions were used in order to compute Figure 4.

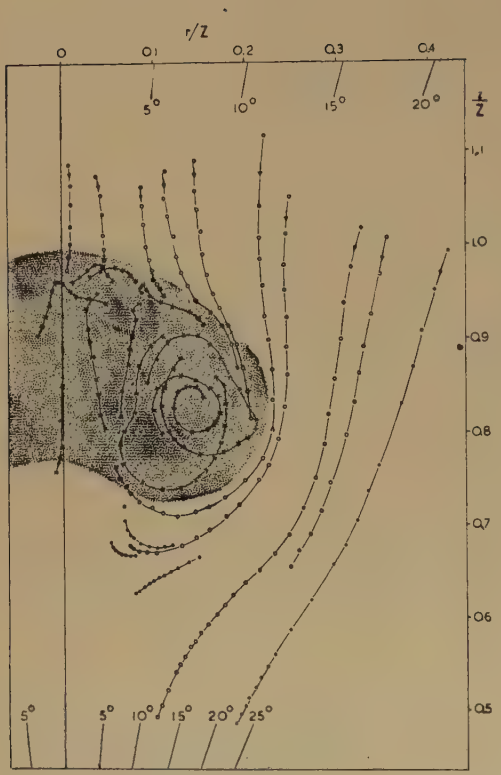


Figure 3. Successive positions of particles obtained from actual measurements of two experiments. The area of buoyant fluid is shaded. The flow is relatively smooth except at the leading edge of the thermal.

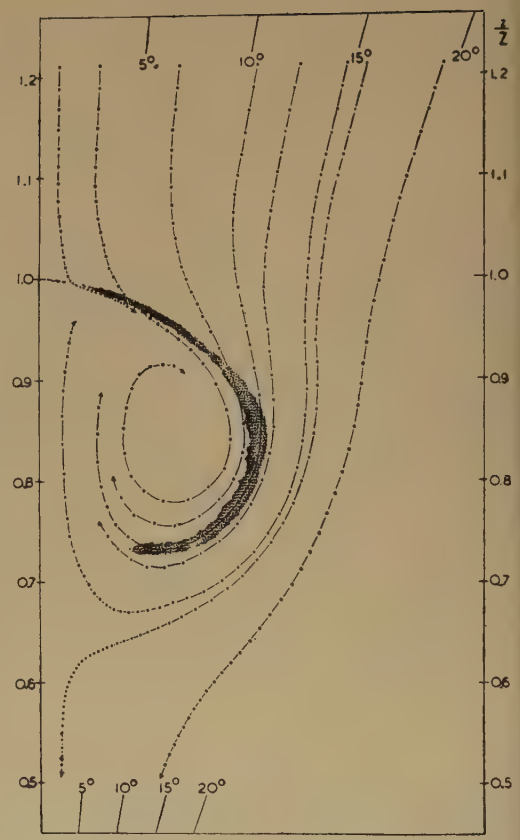


Figure 4. Paths of particles of fluid relative to the thermal and computed from the average velocities presented in Fig. 2.

With this system of co-ordinates, in the absence of the thermal all fluid particles would converge along radii towards the virtual origin. Therefore the outline of the thermal remains fixed and particles follow the tracks shown.

It can be seen that if the thermal grows along the cone of about 15° then all the particles lying inside this cone will eventually enter the thermal because the two streamlines originating on either side of 15° part company at the rear. Some particles will enter at the front, others at the rear.

Approximately 50 to 70 per cent of the particles which enter the thermal will enter it in the front, that is, before the maximum radius is reached. The remaining particles will enter from the rear after they have been pushed to the side by the outflow, brought towards the centre by the inflow and into the region where the vertical velocity is great enough to bring them into the thermal.

This variation from about 50 to 70 per cent depends upon whether the particle is captured by a protuberance. The particle in Fig. 4 which is initially following the angle of about 10·5° will probably enter the thermal before the maximum radius is reached. The particle at 12·0 (or 12·5) will enter at the rear unless engulfed at the side by a protuberance.

Because of the entrainment of the fluid at the rear there is no obvious way of defining the rear boundary of the thermal. In the diagrams it is represented by the approximate position of the shaded (buoyant) fluid. The particle entering from the outside will

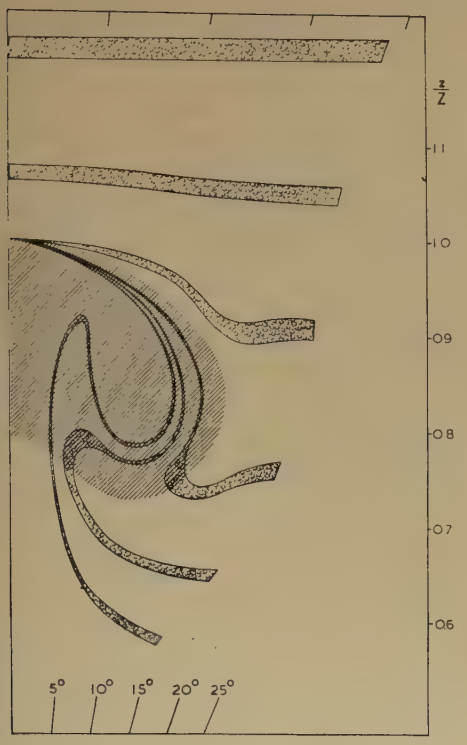


Figure 5. The distortion of an initially horizontal layer of fluid which is caused by the mean motions of the thermal.

Figure 6. The lines show the boundaries where heavy particles will just remain in, or just fall out of, the thermal.

generally not be mixed with the more buoyant particles inside until it has advanced towards the cap. In the laboratory experiments the thermal was ' hollowed-out ' when viewed from the rear.

Fig. 5 was obtained from Fig. 4 and shows the distortion of an initially horizontal layer of fluid which is caused by the mean motions of the thermal; no account is taken of the turbulent or mixing motions.

6. Fallout particles

The paths shown in Figs. 3 and 4 are those of suspended particles, that is, particles which have a zero terminal velocity in the fluid. The tracks of heavy particles were then considered to see what sinking velocity they could have without falling out of the thermal. The results are shown in Fig. 6.

Particles, inside the line marked 0·4, with a terminal velocity of less than 0·4 W_c, where W_c is the rate of rise of the cap, will not escape from the thermal but will be carried towards the axis by the inflow at the rear and up the centre again.

Of the particles with a terminal velocity of 1·0 W_c only those situated to the left of the line marked 1·0 will be retained; those outside this line will fall out of the bottom of the thermal.

The greatest terminal velocity which any particle can have and still remain in the thermal is 1·6 W_c and these particles must be poised on the axis. (In practice, the terminal velocity must be considerably less than 1·6 because of the fluctuations in the centre).

Because of the inflow at the rear, particles which just fall out will fall down the axis. Because the thermal is decelerating, all particles with non-zero terminal velocity will eventually fall out.

7. Comparison with the atmosphere

It has been argued that the isolated thermal is the ultimate form of all convective elements in the atmosphere. Regardless of the method of release or the shape of a convective element in the lower levels it will eventually take the model form presented above, when rising through neutral surroundings.

From flight observations by the author and other glider pilots it appears that this ultimate form is reached at an altitude of about 500 m in countries such as England.

The effects of the motion in and around isolated thermals on a circling sailplane have been described by Woodward (1958). The effects can be summarized as follows :

(a) The sailplane will rise relative to the thermal cap until it reaches a height of about 0·8 or 0·9 times the height of the cap above the point source and it will then continue to rise at about 0·8 or 0·9 times the rate of rise of the thermal cap. (The relative position reached by the sailplane is of course dependent upon the radius of the thermal, the vertical velocity of the cap, and the turning radius, sinking speed and angle of bank of the glider); (b) if the glider is not spiraling concentrically in the field of inflow or outflow (which is usually the case in practice) then the inflow in the lower part of the thermal will tend to centre it; in the upper portion the outflow will tend to ' throw it out,'; (c) as pointed out by Goodhart (1958) the radial flow produces an increment, s, in the vertical velocity of the glider which is dependent on the radial velocity, U_r and ϕ, the angle of bank of the glider. When concentric in the radial flow $s = U_r \tan \phi$; (d) because of the radial velocities, the sailplane will encounter an increase in horizontal velocity when entering (in level flight) the upper portion of the thermal and a decrease when entering the lower.

The pilot, of course, does not know his vertical position in the thermal. However, on numerous occasions it is found that shortly after entering the thermal, relatively strong vertical velocities are encountered, it is advantageous to increase the angle of bank of the glider and one is able to ' centre the thermal ' with comparative ease. After a few minutes the angle of bank may be decreased and strong vertical velocities are still encountered. Several minutes later the vertical velocity gradually decreases, it is advantageous to decrease the angle of bank and it is difficult to ' centre the thermal.'

These observations verify the predicted theoretical effects presented above in (a), (b) and (c) and indicate that the glider has entered the lower portion of the thermal and then climbed relative to the cap.

When entering or leaving a thermal, while in level flight, an increase or decrease in airspeed is sometimes observed. This airspeed change cannot be fully accounted for by the vertical velocities alone and the discrepancy may be due to the radial flow.

The isolated thermal model can also help to interpret an additional flight observation. There have been numerous reports by glider pilots that, when flying at, and just above, cloud base, the cloud often appeared to be ' hollow ' in the centre, i.e. the base of the centre of the cloud could be a hundred metres or so above the base on the edges. Because of the entrainment at the rear of the isolated thermal, the centre of the cloud should be ' hollow ' if the surrounding air has a lower relative humidity.

It is important to note that no attempt is made to show similarity between this model thermal and convection in the lower several hundred metres of the atmosphere. It is extremely doubtful that the thermal is immediately detached from the ground. Considering that the isolated thermal will follow a cone of half angle about 15°, the theoretical radius at heights between 1,000 and 2,000 m would be about 1·5 times that observed. I suggest that the thermal leaves the ground as a ' column ' or ' intermittent plume.' Laboratory experiments have shown that the buoyant plume has a cone of half angle about 11°. The plume will eventually become detached from the ground and because

there is more mixing with the surrounding air at the cap than at the rear, the latter will rise relative to the cap until it forms an isolated thermal. Observations of laboratory models have shown this to be the case but to date no quantitative measurements have been made of actual velocities.

The column may have rotation about a vertical axis while still attached to the ground and if so, this rotation will increase while the column is rising through the super-adiabatic layer. Above the unstable layer the column, or plume, will follow a cone of half angle 8-12° and once it becomes detached from the ground the rotation about a vertical axis will gradually decrease because the particles are moving away from the axis and surrounding non-rotating air is mixed in. The column will start to become an isolated thermal and the cone angle will gradually increase to about 15°. Apart from desert regions the ultimate form will probably be reached between 200 and 2,000 m.

Flight observations show that the isolated thermal model is more pronounced in anticyclonic conditions. In the seconditions those clouds that form have a relatively small diameter and depth and last from about 5 to 20 minutes. When larger cumulus persist, a larger-scale motion, also similar to a ring vortex, is probably set up and the motion of the isolated thermal is superimposed upon this. The scale of the individual thermal is between about 100 and 1,000 m; the scale of the larger motion may be from about 2 to 20 km.

Acknowledgments

The author is indebted to The Munitalp Foundation Inc. for its support; to Professor P. A. Sheppard for providing facilities for work in the Department of Meteorology, Imperial College; and to Dr. R. S. Scorer and Mr. P. M. Saunders for their assistance, collaboration and many helpful suggestions.

References

Goodhart, H. C. N.	1958	*Swiss Aero. Revue*, OSTIV Section, **6**, p. 324.
Scorer, R. S.	1956	*Ibid.*, **9**, p. 432.
	1957	*J. Fluid Mech.*, **2**, 6, p. 583.
	1958	*Natural aerodynamics*, Ch. VII, Pergamon Press, London.
Scorer, R. S. and Ronne, C.	1956	*Weather*, **11**, p. 151.
Woodward, B.	1958	*Swiss Aero. Revue*, OSTIV Section, **6**, p. 321.

551.573

Some observations of potential evaporation, 1955-1957

By F. H. W. GREEN
Nature Conservancy, London

(Manuscript received 10 March 1958; in revised form 3 Dec. 1958)

SUMMARY

The methods of interpreting routine observations with an apparatus designed to measure potential evaporation (PE) are discussed, and tables of results are presented. With a view to standardizing PE measurements, conditions are defined, resembling those for standard climatological stations. Difficulties in the interpretation of daily observations, because of various kinds of time-lag, and variations in storage of water in the apparatus, have been encountered but largely overcome. The values obtained are compared with those computed from other available data by the methods of Penman, Thornthwaite, and Crowe, in which the conditions are either specified, or tacitly assumed, to be similar. The values are in general agreement, but it is shown that at west coast stations in Britain the observed values tend to be lower than the computed, and, more strikingly, that at high-altitude stations, they tend to be higher.

1. INTRODUCTION

Measurements of evaporation from a vegetation-covered surface can be of value in agriculture and hydrology, as well as in soil and natural vegetation conservation. When the water supply is maintained at a high level the quantity then measured is known as the potential evaporation (E_0 or PE) and, if the technique of measuring it can be standardized, PE can be an important and useful climatic discriminant, to be related to other weather elements that determine it. Attempts to establish such a relation already exist (Penman 1948, Thornthwaite 1948, Crowe 1957). Experience shows that any standardization must provide homogeneity of site over a large area round the test equipment, and for all measurements reported here the aim was a sward consisting of the same plant community as that on the apparatus for at least 50 ft in most directions, and vegetation of similar physique, if not the same kind, for a much greater distance. A flat site is desirable, but is not always attainable, and Moor House, Prabost and Coalburn installations are on slopes. Some details of sites appear in the accompanying tables.

2. APPARATUS AND TECHNIQUE

The apparatus and technique have previously been described (Garnier and Lewis 1954, Green 1956, Green 1958a). A water-tight oil drum (Fig. 1) fitted with bottom outlet to a receiver for percolate, is set in the ground, filled with a fairly light soil and turfed with the sods removed to start the excavation. At 09 GMT daily (15 GMT at Coalburn) measurements

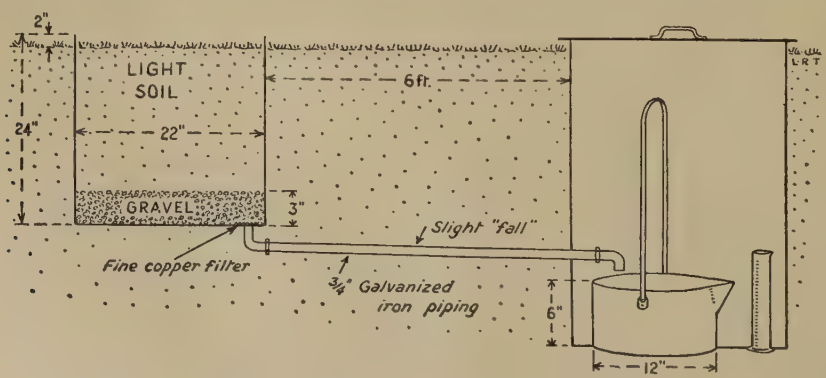


Figure 1. Simple evaporation gauge.

are made of the rainfall for the past 24 hr (R), and percolate for the past 24 hr (N). Except after heavy rain a measured quantity of water (A) is then sprinkled on the surface of the tank, usually the equivalent of 0·10 in. of rainfall in winter and 0·35 in. in summer. Evidence from Coalburn, and a few experiments elsewhere, suggest that the time of watering is not important.

3. Interpretation of observations

For the conditions imposed, the water balance equation of the system is

$$E_0 = (R + A) - (N + S)$$

where E_0 is the evaporation and S in the *increase* in stored water in the soil column. As S cannot satisfactorily be measured, the technique is designed to keep it as near zero as possible by frequent watering, and noting dates when it is actually zero. It is known that about 90 per cent of the water excess percolates in 24 hr, and it is a reasonable presumption that after 72 hr the amount still to drain is negligible. Accordingly, if the same amount A is added each day, S can be assumed to be zero for a period of three days without rain, and then $E_0 = A - N$, as long as N is positive. If N is kept small, the soil is then maintained close to field capacity, and except when rainfall is heavy and frequent, a skilled operator can maintain this condition by matching the value of A with known R. The method has been found to break down only when the rainfall is so heavy and frequent, and at such irregular intervals, that it becomes impossible to have recognisable times of equal S within the period required; at Prabost or Kinlochewe in winter, for example, it is possible for more than a month thus to go by without such time points being discernible.

One example given in Fig. 2 shows how the above procedure normally works out in practice. The low points in the continuous line showing cumulative E_0 are points where values of S are nearly identical, and in this case approximately at field capacity. Thus, for instance, over the period 1 to 18 Aug., E_0 is seen to be 1·35 in. with a margin of error of less than 0·10 in. i.e., less than an average of 0·005 in./day. Such small margins of error are the less important in that they are cancelled out over a longer period, and lead only to a small carry-over from one unit period to another. If one works in calendar months, the drawing of a smooth line between the ' time points ' of equal storage (or when S is zero)

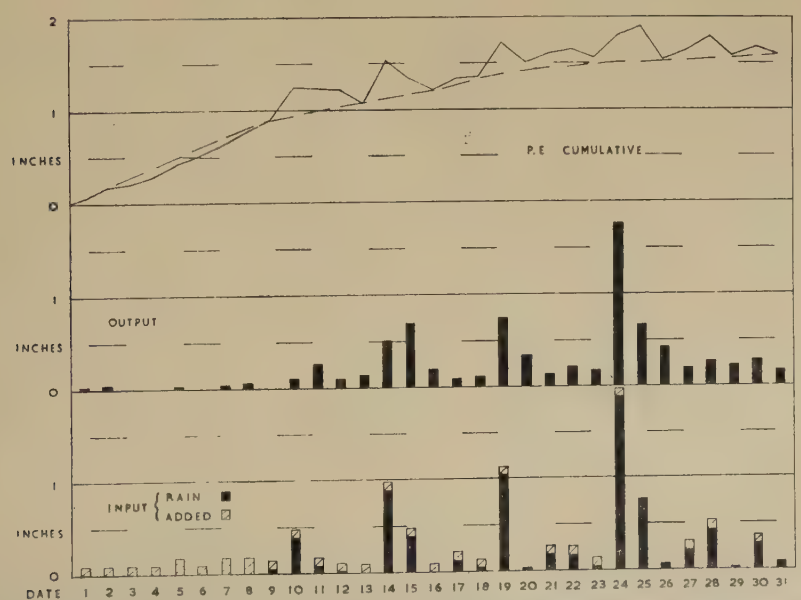


Figure 2. Prabost, Isle of Skye, August 1957. Showing the effects of (a) insufficient water added to maintain field capacity, 3 to 9 August, and (b) surplus of water in tank at intervals after this.

TABLE 2. POTENTIAL EVAPORATION (Values are for calendar months ending 09 GMT on the last day of the month. Values in inches. Figures in brackets are uncertain).

The computed figures have been obtained by Crowe's method but without making adjustment for rainfall in the warmest month. This formula is much easier to use than Penman's or Thornthwaite's, and appears to give values *at least* as close, on the average, to the observed figures, as either of the others. Computed values, by the latter, are however, given for Wellesbourne.

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1955			ACHNAGOICHAN c. 1,000 ft. O.D. (57°09 N, 3°19 W) Nat. Grid Ref. 28/913082.								(Observer: A. G. MacDonald)		
PE observed	(1·30)	(0·30)	0·79	2·58	(1·98)	(2·72)	3·50	4·40	3·62	1·07	0·72	0·37	23·35
PE computed	0·25	0·55	0·88	2·39	2·66	3·57	4·64	4·11	1·93	0·93	0·32	0·24	22·47
Rainfall	1·33	(4·10)	1·14	0·70	3·16	0·40	0·84	1·43	2·20	3·01	0·95	4·30	23·56
1956													
PE observed	0·25	0·00	1·50	1·38	(3·35)	(3·85)	3·66	1·77	1·78	0·72	0·47	0·50	19·23
PE computed	0·23	0·45	1·02	1·65	2·19	2·81	3·20	1·99	1·90	0·83	0·29	0·16	16·72
Rainfall	1·98	2·48	1·42	1·93	1·30	3·20	4·47	6·26	4·23	3·30	1·58	4·09	36·24
1957													
PE observed	0·30	− 0·20	1·04	1·91	2·63	3·52	1·95	1·73	1·20	0·77	0·40	0·05	15·30
PE computed	0·20	0·35	0·91	1·40	2·39	3·20	2·29	2·11	1·23	0·69	0·22	0·14	15·13
Rainfall	3·14	2·88	2·58	1·15	1·36	2·44	4·63	4·84	2·25	2·01	3·37	2·77	33·42
1956			LOSSIEMOUTH c. 15 ft. O.D. (57°42 N, 3°19 W) National Grid Ref. 38/2170. (Senior Meteorological Officer, R.N. Air Station)										
PE observed	–	–	1·31	2·02	(3·90)	2·85	2·85	3·49	1·81	1·32	0·38	(0·17)	–
PE computed*	0·18	0·42	1·21	1·63	3·56	2·55	3·36	2·01	1·76	0·92	0·30	0·13	17·03
Rainfall	(2·30)	(0·80)	0·66	1·52	1·22	2·36	4·99	5·01	2·03	2·39	1·38	1·95	26·61
1957					* for Kinloss (about 10 mi WSW)								
PE observed	0·68	0·53	0·76	(2·86)	(2·38)	3·40	2·78	2·78	2·56	1·37	0·22	0·38	20·70
PE computed*	0·35	0·60	1·64	1·36	3·24	4·47	3·04	3·17	1·76	1·26	0·48	0·22	21·59
Rainfall	2·30	0·80	2·51	0·46	0·77	1·50	5·02	3·49	1·14	1·03	1·53	1·42	21·97
						* for Kinloss							
1956			KINLOCHEWE c. 80ft. O.D. (57°36 N, 5°19 W) Nat. Grid Ref. 28/024630.								(Observer: J. Polson)		
PE observed	–	–	(1·30)	(1·33)	(1·24)	2·32	2·61	2·17	1·46	(− 0·35)	(0·38)	(0·15)	12·61*
PE computed	–	–	1·53	2·01	2·25	3·54	3·39	1·56	1·93	1·00	0·37	0·17	17·74*
Rainfall	8·58	5·99	4·32	5·48	7·23	3·12	5·59	6·04	3·83	11·22	10·32	14·14	85·86
If 5% added to monthly rainfall													
PE observed	–	–	1·52	1·60	1·50	2·48	2·89	2·47	1·65	0·21	0·90	0·86	16·08*
Rainfall	9·01	6·30	4·54	5·75	7·49	3·28	5·87	6·34	4·02	11·78	10·84	14·85	90·07
1957						* for 10 months							
PE observed	− 0·25	− 0·35	0·29	1·52	1·90	2·97	2·44	1·79	0·92	0·89	− 0·08	0·22	12·26
PE computed	0·27	0·46	0·97	2·11	2·77	3·28	2·38	2·56	0·88	1·04	0·29	0·16	17·17
Rainfall	14·91	5·44	5·78	3·13	3·24	3·16	4·76	7·97	6·58	11·11	5·26	13·80	85·14
If 5% added to monthly rainfall													
PE observed	0·50	− 0·08	0·58	1·68	2·06	3·13	2·68	2·19	1·25	1·60	0·34	0·90	16·83
Rainfall	15·65	5·71	6·07	3·29	3·40	3·32	5·00	8·37	6·70	11·67	5·52	14·49	89·19

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1957													
PE observed	(− 1·85)	(− 0·28)	1·47	1·93	2·37	3·31	2·63	1·62	0·68	0·32	− 0·10	− 0·19	11·91
PE computed*	0·38	0·67	1·53	2·09	2·76	4·01	3·01	2·64	1·97	1·22	0·39	0·20	20·87
Rainfall	9·25	5·02	6·75	3·98	2·71	3·07	3·49	7·16	5·19	11·84	5·41	9·97	73·84
If 5% added to monthly rainfall													
PE observed	− 1·39	0·17	1·81	2·13	2·51	3·46	2·80	1·98	0·94	0·91	0·17	0·31	15·80
Rainfall	9·71	5·47	7·09	4·18	2·85	3·22	3·66	7·52	5·45	12·43	5·68	10·47	77·73

Prabost c. 200 ft. O.D. (57°29 N. 6°17 W) Nat. Grid Ref. 18/4285. (Observers : Mr. and Mrs. R. M. Murray)

* for Duntulm, Isle of Skye (about 16 mi to north)

1957													
PE observed	0·68	0·85	0·84	1·87	3·26	(4·60)	(3·45)	2·98	1·48	0·78	0·32	0·56	21·67
PE computed	0·34	0·62	1·27	1·90	3·83	4·61	3·25	2·75	1·60	1·05	0·36	0·24	21·82
Rainfall	6·39	3·45	4·83	2·34	2·69	2·53	4·01	6·27	3·76	5·47	3·02	5·58	50·34

Coalburn c. 930 ft. O.D. (55°35 N Long. 3°55 W) Nat. Grid Ref. 26/8034. (Observer : Capt. J. Ross)

1957													
PE observed	(0·80)	(1·00)	(1·28)	1·65	(1·44)	3·60	3·36	(3·20)	0·86	(0·90)	(0·42)	(0·75)	19·02
PE computed	0·34	0·37	1·03	1·63	2·62	4·34	2·62	2·30	1·25	1·30	0·31	0·27	18·38
Rainfall	9·78	9·56	5·93	1·41	3·74	2·13	8·10	10·34	6·73	7·12	4·32	8·24	77·40

Moor House c. 1,830 ft. O.D. (54°41 N. 2°23 W) Nat. Grid Ref. 35/757328. (Observer : M. Rawes)

1957													
PE observed	− 0·72	− 0·37	0·68	2·28	3·47	4·56	2·10	2·12	(1·25)	(0·37)	(− 0·46)	− 0·20	15·08
PE computed	0·34	0·56	1·10	1·48	2·38	3·78	2·22	2·12	1·29	1·00	0·45	0·35	17·07
Rainfall	3·09	4·79	4·11	0·49	1·77	3·40	7·11	6·79	8·91	5·73	4·35	3·91	54·45
If 5% is added to monthly rainfall													
PE observed	− 0·57	− 0·13	0·94	2·30	3·56	4·63	2·46	2·46	1·70	0·66	− 0·24	0·00	17·77
Rainfall	3·24	5·03	4·37	0·51	1·86	3·57	7·47	7·13	9·36	6·02	4·57	4·11	57·24

Swyddffynnon c. 570 ft. O.D. (52°16 N. 3°35 W) Nat. Grid Ref. 22/696655. (Observers : Mr. and Mrs. R. M. Kydd)

(The following observations, made with similar apparatus, are given with acknowledgments to the authorities named)

1957													
PE observed	0·15	0·11	0·84	1·39	3·38	3·40	2·19	2·57	1·58	0·98	0·31	0·21	16·91
PE computed	0·45	0·67	0·95	1·88	2·43	3·58	2·39	2·15	1·24	0·96	0·60	0·36	17·66
Rainfall	7·00	5·77	7·78	2·74	2·82	1·85	5·63	3·34	6·17	5·72	2·67	5·00	56·59

Valentia, Co. Kerry, Ireland (Irish Meteorological Service).

1957													
PE observed*	0·94	0·41	1·22	2·00	3·08	5·51	3·18	3·00	1·59	0·79	0·39	0·39	22·50
PE computed	0·37	0·60	1·32	1·84	2·57	4·64	3·09	2·75	1·68	1·31	0·45	0·32	20·94
Rainfall*	1·67	2·38	1·97	0·22	1·34	1·03	2·70	2·64	3·13	1·23	1·51	2·31	22·13
PE Penman	0·22	0·20	1·00	1·74	3·57	4·18	3·04	2·47	1·29	0·64	0·18	0·13	18·74
PE Thornthwaite	0·67	0·71	1·81	2·04	2·67	4·07	4·53	3·74	2·59	1·85	0·84	0·50	26·02

Wellesbourne, Warwicks. (National Vegetable Research Station).

* Except for days of snow, precipitation was measured by raingauge at 0·8 in. above ground level surrounded by fibre mat to prevent insplash; this collected through the year 105 per cent of the amount measured at the standard height of 1 ft.

will enable the amount for the whole month to be read off with reasonable accuracy; this is illustrated in Fig. 2 by reading off E_0 from the graph for the first 15 days of August as 1·16 in.

During the periods when the shortage of water in the area surrounding the apparatus is acute, an ' oasis ' effect has to be taken into account. It is found that this can be allowed for by applying the results of a study by Sutton (1934) as described by Green (1958b). Briefly, it can be stated that the total evaporation from a circular wet surface in completely dry surroundings, increases not according to the square radius, but according to the 1·88 power of the radius. This conclusion, although reflection shows that it can apply only within certain limits of areal dimension, appears in practice to be applicable within the limits of 11 in. (the diameter of the drum) and 200 yd radius; it is in fact applicable to a much greater distance than 200 yd from the centre, but increasing the radius beyond this limit alters the reduction factor only negligibly.

Some observations made with the apparatus at two sites in Malta in 1957 clearly illustrate the application of this reduction factor. At Hal-Far there is no irrigation of the surrounding area, but at Ta-Qali there is a surrounding irrigated area approximately 20 ft in radius. ' Reduction ' of the observed values of PE to the equivalent of those which would be expected on an irrigated area of 200 yd radius is shown in Table 1*.

TABLE 1.

	Hal-Far		Ta-Qali	
1957	Observed (inches)	reduced (inches)	observed (inches)	reduced (inches)
June	11·82	5·42	7·36	5·21
July	13·90	6·38	9·82	6·51
August	18·05	8·31	12·04	8·00

1956 and 1957 had fairly wet summers and, in fact, the reduction factor has been applied only to the Achnagoichan figures for June to September 1955. The brackets placed round some of the figures in the tables are, in respect of the summer months, due to uncertainty as to whether the reduction factor should not have been applied in those cases also. It almost certainly should have been applied to Achnagoichan in the early summer of 1956 and Coalburn in the summer of 1957; it will immediately be seen that such reduction would bring the values more into line with the computed ones.

Table 2 gives the observed figures for a calendar month, after adjustment for time-lag, temporary over-exposure, and correction of apparently wrong entries. (Values in brackets are doubtful for one or other of these reasons, or because the precipitation was uncertain.) Certain anomalies still remain, some of which are features of individual stations. There are some anomalies, however, which are more or less general :

(i) the raingauges appear to have been over exposed at some stations, as noted by Stanhill (1958) at Wellesbourne. If Stanhill's estimate of a 5 per cent minus error in reading is accepted, the PE values can be adjusted as shown in the tables, and it will be seen that at wet, west coast sites the values are then in closer agreement with computed values for those sites. Fig. 2 illustrates a case where PE may have been underestimated for this reason, in the period commencing 20 Aug. 1957. This matter has been referred to in a recent note (Green 1958), and is being studied further.

(ii) Snowfall aggravates the above difficulty, so much so that observations on certain days have virtually to be disregarded, and values of PE have been interpolated in such cases.

* I am indebted for these figures to Mr. P. K. Mitchell, University of Durham, who arranged, after discussion with me, to start the observations, by courtesy of the Naval authorities (at R.N. Air Station, Hal-Far), and the Senior Water Engineer, Government of Malta (at Ta-Qali).

(iii) There is some evidence that ' field capacity ' of the soil tanks, especially in the first stages of operation with disturbed soil, has some seasonal variation, becoming less in winter and increasing in spring. This would give a long-period time-lag, which is self-compensating over a twelve months period.

(iv) There still remains a tendency for observed value of PE at wet, west coast stations to be significantly lower than computed values, and for those at high-altitude stations to be higher. This suggests that there are elements in the climatic environment which are not taken adequately into account in the formulae.

4. Discussion of results

While it is clear that the variations in incoming solar radiation, corresponding with latitude and season, and usually expressed by mean maximum temperature and length of day, account for by far the greater part of the seasonal variations in PE, the effects of the changing synoptic situation can be detected as ripples on the main curve.

There are considerable variations in the total annual PE at any one place from one year to another. For example, Achnagoichan varied from about 15½ in. in 1957 to about 23½ in. in 1955; and adjustment for underestimation of rainfall could effect this range only very slightly.

As between different places in one year, systematic regional differences in PE are clearly shown, and these agree in general with those revealed by values computed by Crowe's method (1957). If adjustment (by adding 5 per cent of total annual rainfall) is made where it seems appropriate to do so, because of evidence that the raingauge is over exposed, the 1957 values, in order of magnitude, are as follows :

	(inches)
Wellesbourne	22·8
Lossiemouth	21·8
Coalburn	21·7
Moor House	19·0
Swyddffynnon	17·8
Achnagoichan	17·0
Valentia	16·9
Kinlochewe	16·8
Prabost	15·8

There is clearly a general fall from south to north and from east to west, but elevated stations all appear higher in the list than their latitude and longitude alone would suggest. There is some indication that this anomaly is due primarily to maintenance of higher PE rates in winter than at lowland sites, and this is consistent with the fact that there is less difference in accumulated temperature between winter and summer months at higher elevations than at lower ones.

5. Conclusions

It will be apparent from the above account that certain improvements need to be made in the installations at various stations, but it is also apparent that observations with relatively cheap and simple apparatus give quite reasonable results, and indeed much less crude than expected, if the site conforms approximately to the conditions normally specified for climatological observing stations. Attention needs to be given particularly, however, to rainfall recording, especially in regions of heavy precipitation, and this is now being done at Achnagoichan and Kinlochewe, and at a new station in the Isle of Rhum.

The many lessons to be learned from a network of PE stations and the values of this element in relation to the other factors, and to the soil and vegetation, are not discussed here, but are being examined, and it is hoped soon to make some of them the subject of a separate paper.

Acknowledgments

I am indebted to many people for assistance in this work, notably the observers named in the Tables, and my colleague Mr. J. G. Skellam, who has advised me frequently on statistical method. I am also indebted to Mr. J. H. L. Sibbons, of Sheffield University, for drawing my attention to the relevance of the paper by Sutton (1934), and also to Dr. J. R. Mather, of the Laboratory of Climatology, Elmer, New Jersey, for very helpful comments on operational technique in the early stages.

References

Crowe, P. R. 1957 *Geographical Studies*, **5**, 1, pp. 56-75.
Garnier, B. J. and Lewis, W. V. 1954 *Weather*, **9**, 8, pp. 243-245.
Green, F. H. W. 1956 *J. Inst. Water Eng.*, **10**, 5, pp. 411-419.
1958a *Weather*, **13**, 8, pp. 275-277.
1958b *Transactions 1957 Toronto Assembly International Assoc. Sci. Hydrol.*, Gentbrugge, Vol. III, pp. 444-451.
Penman, H. L. 1948 *Proc. Roy. Soc.* A, **193**, pp. 120-145.
Stanhill, G. 1958 *Weather*, **13**, 1, pp. 33-34.
Sutton, O. G. 1934 *Proc. Roy. Soc.* A, **146**, pp. 701-722.
Thornthwaite, C. W. 1948 *Geograph. Rev.*, **38**, p. 56.

551.551.21 : 551.556.8

Wind stress on a water surface : measurements at low wind speeds with the aid of surface films

By R. G. VINES

Division of Physical Chemistry, Commonwealth Scientific and Industrial Research Organization, Melbourne, Australia

(Manuscript received 5 September 1958; in revised form 19 January 1959)

SUMMARY

A monolayer on a water surface is affected by the action of wind, and retracts until the wind drag is counter-balanced by the opposing surface pressure gradient. This effect has been used to determine the shearing stress on water at low wind speeds. Estimates of the drag coefficient have been obtained.

1. INTRODUCTION

Many attempts have been made to determine wind stresses on water, but, despite experiments at sea (Deacon, Sheppard and Webb 1956), in wind tunnels (Francis 1951, Keulegan 1951) and on small lakes (Van Dorn 1953), much uncertainty still remains. At very low wind speeds accurate measurement is difficult and no reliable determinations have hitherto been reported.

When an insoluble monolayer is spread on water, a surface pressure is established. In a study of the influence of wind on surface films, it was noticed that the extent of a film of known surface pressure varied markedly with wind velocity; the present note shows how this effect can be used to measure wind stresses at low wind speeds. Experiments using detergent films have previously been described by Keulegan (1951) and Van Dorn (1953). However, in these experiments, detergent was used merely to calm the water surface, and the influence of wind on the films themselves was not investigated.

2. EXPERIMENTAL METHOD

The apparatus consisted of a metal trough 160 cm long, 12 cm across and 4 cm deep; it was completely filled with water, and was housed in a small tunnel provided with an electric fan of variable speed and an adjacent wire-mesh screen to help smooth out the airflow. The working length of the tunnel was about $7\frac{1}{2}$ ft and it was roughly one foot square in section. A small wooden fairing, the same width as the tunnel, was placed against the trough at the windward end. This served to reduce turbulence at the trough wall and lead the airstream gradually up to the water surface.

At the far end of the trough was a sensitive film balance (Harkins and Anderson 1937) which was used to measure the surface tension of the water. The unit consisted of a thin mica plate roughened with emery paper (Matalon and Schulman 1949) and supported vertically by a fine quartz spiral. The mica sheet was held parallel to the airstream so that very little force was exerted upon it by air movement.

When surface-active materials were introduced to the water in the absence of wind, a monolayer was observed to spread over the entire area of the trough. Wind, however, caused retraction of the film, until the wind stress was exactly counter-balanced by the opposing film pressure gradient. Since the surface pressure of a monolayer is equal to the difference between its surface tension and that of clean water, the surface pressures set up at the end of the trough by the action of the wind were readily obtained from the balance readings.

In the absence of surface films the wind stress was compensated by a slight tilting of the water surface, and the small differences in level so produced were manifest by the return movement of water at the bottom of the trough. If this movement was restricted by a baffle, water could be made to flow back through a narrow tube joining both ends; the flow was easily visible if an air bubble was trapped in the tube. When a film was present, such circulation of water came to an end, indicating that the wind stress was compensated entirely by the pressure gradient of the film, and no energy spent in bulk movement of fluid. Confirmation of this was obtained by introducing solid fragments of potassium permanganate into the water : only in those regions where the surface was not covered by a film was any bulk circulation detected.

The materials employed were tricresyl phosphate, ethyl palmitate and lauric acid. As these substances were not of high grade, pressures greater than the accepted equilibrium pressures were often established, owing to the presence of impurities : the results obtained were dependent only upon the film pressures actually set up, and not upon the materials used.

Wind speeds at a height of ~ 5 cm above the water surface were measured with a Simmons shielded hot-wire anemometer (N.P.L. calibration), and with an Alnor 'velometer'* results obtained by both methods agreed to within about 10 per cent. The distance a surface film extended along the trough was observed by dusting a small quantity of talc or powdered chalk on to the water; this was blown freely over the uncovered surface, but remained stationary for a time at the film front. Measurements were made of the extent of the film and the corresponding film pressure exerted at the balance. Additional experiments were carried out using a second trough which was narrower, shorter and deeper than the one originally employed, but no significant differences in the results were observed, as is to be expected from the results of experiments on bulk movement beneath the water surface.

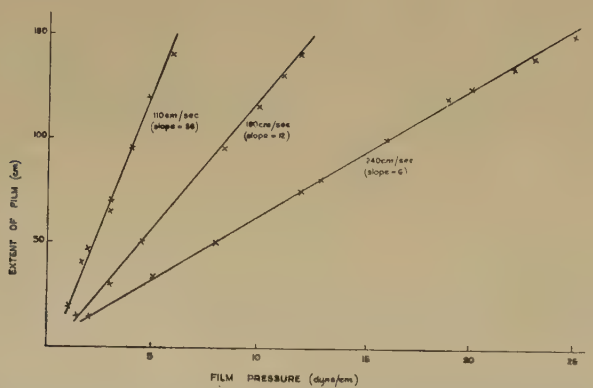


Figure 1. The determination of wind stresses : the relationship between surface pressure and extent of film, at various wind speeds. The slopes shown are reciprocal surface-pressure gradients (units cm^2 $dyne^{-1}$).

3. Results

The results obtained at wind speeds of 110, 160 and 240 cm sec^{-1} are shown in Fig. 1. At these wind speeds there is a linear relationship between the extent of the film and the the film pressure at the end of the trough. This implies that there is a linear variation in pressure up to the film front, and measurement of pressures along the trough confirmed this view. There appeared to be a small pressure discontinuity at the film front itself, but it amounted to only 1 dyne cm^{-1} or less : for this reason the lines in Fig. 1 are not directed exactly towards the origin.

* This instrument was connected to a small nozzle inside the tunnel, and the main gauge was placed outside so that the flow of air was unaffected. Measurements were obtained at the middle of the trough, and at one end near the pressure balance; in both these positions air velocities at the edges of the trough were almost the same as those at the centre.

Values of τ, the stress per cm^2 of surface, are given in Table 1 together with corresponding values of the drag coefficient C as calculated from the relation

$$C_h = \tau/\rho\, u_h^2 :$$

here, ρ is the density of air (0·0012 g cm^{-3}), and u_h is the wind velocity at a height h metres above the surface.

TABLE 1. DRAG COEFFICIENTS AT LOW WIND SPEEDS

$u_{0\cdot05}$ (cm sec^{-1}) Wind speed at 5 cm	Reciprocal of surface-pressure gradient	τ (dyne cm^{-2}) Surface stress per unit area	$C_{0\cdot05}$ Drag coefficient (5 cm)
110	26	$0\cdot038_5$	$2\cdot6 \times 10^{-3}$
160	12	$0\cdot083_5$	2·7
240	6	0·165	2·45
290†	4	0·25	2·5

† A single experiment was carried out at a wind speed of ~ 290 cm sec^{-1} : a film of pressure 25 dyne cm^{-1} extended 100 cm from the pressure balance.

4. DISCUSSION

It is well known that for moderate winds on a water surface, the drag coefficient C increases gradually with wind velocity, though C assumes a more or less constant value at high speeds when the surface is rough. At low wind speeds the position is uncertain, and the value of the drag coefficient unknown : C may rise slowly with wind speed from very small values at low velocities, as is implied in the graphs of Francis (1954), or it may possibly have a constant value corresponding to that for an aerodynamically smooth surface.

The present measurements show no significant variation in drag coefficient as the wind speed is increased from 1 to 3 m sec^{-1} : it is therefore interesting to compare the results obtained with those predicted from the formula for steady-state flow over a smooth plate. The formula is (cf. Sutton 1953) :

$$\frac{1}{\sqrt{C}} = \frac{u}{u_*} = \frac{1}{k}\log_e \frac{(9\, u_*\, z)}{(\nu)}$$

where, again, u is the air velocity at height z cm above the surface, u_* is the friction velocity $\left(=\sqrt{\frac{\tau}{\rho}}\right)$, ν is the kinematic viscosity of air and k is von Kármán's constant (of value 0·40). Using values of τ from Table 1, air velocities at a height of 5 cm may be calculated; these are 106, 164, 242 and 305 cm sec^{-1} respectively, in extremely good agreement with the corresponding experimental values, viz. — 110, 160, 240 and 290 cm sec^{-1}. It is clear that, at these wind speeds, water covered with a surface film is aerodynamically smooth.

From the formula above, drag coefficients at different heights may be determined. Corresponding to the value $C_{0\cdot05} \sim 2\cdot5 \times 10^{-3}$ given here, values are obtained for $C_{0\cdot1}$, the drag coefficient at 10 cm ($\sim 2\cdot25 \times 10^{-3}$) and C_{10}, the drag coefficient at 10 cm ($\sim 0\cdot9 \times 10^{-3}$). Since surface films are known to damp capillary waves, drag coefficients for a *clean* water surface may be somewhat greater than is indicated by these figures, especially at higher wind velocities when the smoothness of the surface could be affected. However, at low wind speeds, it seems unlikely that the drag coefficients fall *below* these values*.

* When a film is present the surface drift downwind is decreased, and it might be argued that, since the relative speed between wind and water is less for a clean surface, the drag could well be less also. However, Keulegan (1951) has shown that the ratio of surface to wind velocities is never greater than ~ 0·03. In the present experiments the ratio is probably even less, and any correction resulting from this effect will thus be very small indeed. Further evidence that there is little difference in the wind stress on a clean and contaminated surface is obtained from the observation that the pressure discontinuity at a film front is small (see § 3).

Acknowledgment

The advice of Mr. E. L. Deacon, C.S.I.R.O. Division of Meteorological Physics, is gratefully acknowledged.

References

Francis, J. R. D.	1951	*Proc. Roy. Soc.* A, **206**, p. 387.
	1954	*Quart. J. R. Met. Soc.*, **80**, p. 438.
Deacon, E. L., Sheppard, P. A. and Webb, E. K.	1956	*Austr. J. Phys.*, **9**, p. 511.
Harkins, W. D. and Anderson, T. F.	1937	*J. Amer. Chem. Soc.*, **59**, p. 2189.
Keulegan, G. H.	1951	*J. Res. Bur. Stand.*, **46**, p. 358.
Matalon, R. and Schulman, J. H.	1949	*J. Colloid Sci.*, **4**, p. 89.
Sutton, O. G.	1953	*Micrometeorology*, McGraw-Hill, p. 80.
Van Dorn, W. G.	1953	*J. Mar. Res.*, **12**, p. 249.

551.578.11

The composition of precipitation at Lerwick, Shetland

By B. C. V. ODDIE
Meteorological Office, London

(Manuscript received 8 December 1958)

Summary

It is shown that most of the salts deposited at Lerwick arrive as sea-spray rather than in true precipitation. Mg, K and Na are present in very nearly the same proportions as in sea water : but there is a very great deficiency of chlorine, which, it is argued, can only be explained on Cauer's hypothesis that NaCl in aerosol form is dissociated and that Cl or HCl is released.

1. Introduction

A station for the chemical sampling of air and rainwater has been in operation at Lerwick Observatory, Shetland Islands, since 1 January 1958. It is unique in having a very isolated, island situation. The nearest considerable land mass is Scotland, 200 km to the SSW and the nearest industrial area is over 500 km distant. Moreover, the prevailing winds are from the Atlantic, so that the air should be unusually free from any material originating on land.

The sampling unit is situated about 85 m above m.s.l. on the east side of the mainland of Shetland. The surrounding district is shown in Fig. 1. The only town of any size is Lerwick, 4 km to the NE – a place of about 10,000 inhabitants.

2. Examination of results

When the results are examined the first point that strikes one is the enormous quantity of salts collected with the rainwater. In January 1958, the total was about 36 g/m^2. No other station in the European network has ever approached this figure. Even Lista, which is on the Norwegian coast and is the most exposed Continental station, hardly collects as much in a year.

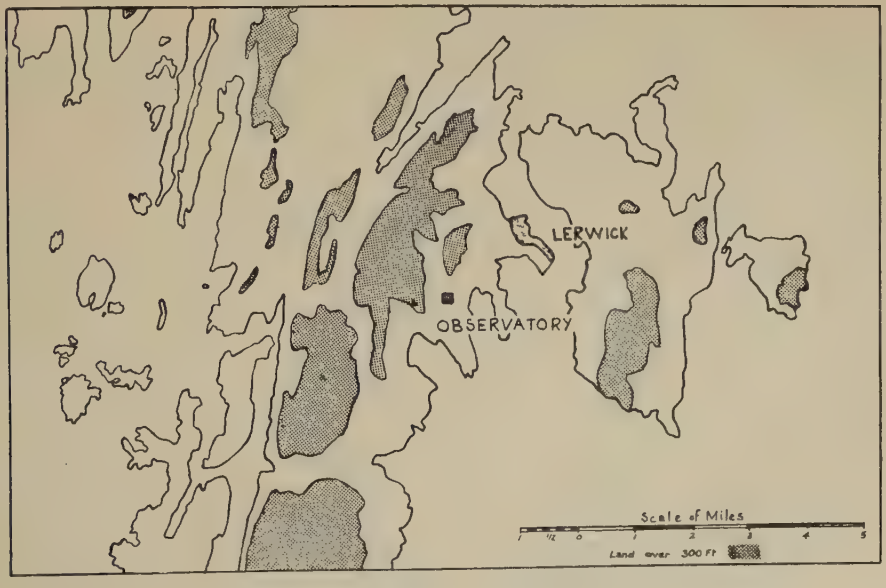


Figure 1. Surroundings of Lerwick Observatory.

Undoubtedly most of this material enters the collecting funnel as sea-spray, which can be clearly seen blowing across the island on any windy day. That this is so is shown by the following consideration – the concentration of salts in the air, as measured by the sampling equipment, is scarcely higher at Lerwick than at an inland station such as, for example, Eskdalemuir. One would expect, therefore, that the same quantity of rain would bring down about the same amount of salt at both stations : in other words, that the salinity of rainwater would be about the same at both places. In fact, it is one or two orders of magnitude greater at Lerwick than at Eskdalemuir It seems likely, therefore, that very large quantities of salts in addition to those brought down by rain are entering the collecting funnel, and this can only be in the form of spray*.

A priori one would expect the collected rainwater to have the constitution of dilute sea water, i.e., that it would contain sodium, potassium, magnesium and chlorine in the same proportions as sea water. This point is investigated in Table 1.

TABLE 1. Composition of rainwater at Lerwick

1958	Mass of Na deposited mg/m²	Ratios of masses deposited × 100 Cl : Na	Mg : Na	K : Na
January	6273	168	12·4	3·7
February	3767	141	12·0	4·1
March	–		No rain	
April	884	166	9·8	4·2
May	1030	169	12·9	3·9
June	148	129	12·2	3·4
July	336	117	9·5	3·3
August	470	152	11·5	4·3
September	522	152	10·5	4·2
	Sea water	182	12·0	3·6

It will be seen that whilst the three metals are present in proportions astonishingly close to the sea-water values, there is a very marked deficiency of chlorine in each month, and that in July this deficiency amounts to no less than 35 per cent.

Low values of the Cl : Na ratio are normal in rainfall analyses, especially at inland stations, and two alternative explanations have been put forward. One, originally advanced by Cauer (1949) and supported by Rossby and Egnér (1955), supposes that NaCl in the form of droplets is oxidized so that chlorine or HCl is released, and is not brought down by rain. It will be seen that the analyses in Table 1 are consistent with this theory, since the constitution of the dissolved material is precisely that of sea salt from which some chlorine has been released.

Junge (1957), on the other hand, has argued that no appreciable dissociation of NaCl takes place, and that low values of the Cl : Na ratio are explicable by sodium salts, other than the chloride, collected by the air in passing over land and subsequently washed out by rain.

It is easy to show that this latter hypothesis cannot account for the Lerwick results. For let [Na] and [Cl] be the amounts of sodium and chlorine brought down/month/m². On Junge's hypothesis, the amount of sodium of marine origin is [Cl]/1·82 (see last line of Table 1) and an amount [Na] – [Cl]/1·82 therefore originates on land. If we calculate this latter quantity for February 1958 (to take a striking case) we find that it is about 74 mg/m² at Eskdalemuir, 10 mg/m² at Aberdeen and Newton Abbot and zero at Leeds, Rothamsted and Edinburgh; whilst at Lerwick it is no less than 850 mg/m². It is obvious that material originating over land could not fall on a remote island in quantities so vastly greater than those which fell on the mainland. Thus the abnormal Na : Cl ratio at Lerwick cannot be due to the presence of sodium from a land source, and must result from dissociation of Na : Cl derived from the sea, as supposed by Cauer.

* Air enters the air-sampling unit through a downward-turned funnel at a speed of about 3 cm sec^{-1}. The object of this device is, of course, to exclude rain; but it also excludes coarse spray, and, in fact, any particle having a terminal velocity greater than 3 cm sec^{-1}.

3. Conclusions

The same conclusion can be reached by a quite independent argument. If a substantial part of the sodium in the Lerwick sample originated on land, it is hard to see why the ratios Mg : Na and K : Na should have so nearly the same values as in sea water. The fact that they have, is, incidentally, quite exceptional, as may be seen from Table 2. This gives the corresponding values for all British stations for August 1958, a month chosen quite at random.

TABLE 2. Composition of rainwater at british stations in August 1958

Stations	Mass of Na deposited mg/m²	Ratio of masses of cations present × 100 Mg : Na	K : Na
Sea water	–	12·0	3·6
Lerwick	470	11·5	4·3
Aberdeen	87	19·5	17·2
Stornoway	307	7·2	3·6
Edinburgh	38	48·4	39·5
Eskdalemuir	417	3·1	10·6
Leeds	28	118·0	46·4
Aldergrove	198	3·3	5·6
Rothamsted	35	28·6	48·6
Camborne	178	14·1	9·6
Newton Abbot	114	21·0	15·8

Acknowledgment

I am indebted to the Director-General of the Meteorological Office for permission to publish this note.

References

Cauer, H. 1949 'Ergebnisse chemisch-meteorologischer Forschung' *Arch. Met. Wien.*, B, **1**, p. 221.

Rossby, C. G. and Egnér, H. 1955 'Chemical climate and its variation with the atmospheric circulation pattern,' *Tellus*, p. 118.

Junge, C. E. 1957 'Atmospheric chemistry,' *Advances in Geophysics*, **4**, p. 92.

551.542.1 : 551.588.5

The pressure gradient at the edge of an ice sheet

By R. A. HAMILTON
Meteorological Office, Prestwick

(Manuscript received 23 October 1958)

SUMMARY

Mean sea-level pressures on the NE coast of Greenland are compared with those of a low-level station within the edge of the inland ice. The comparison indicates that there is no evidence of an unusually high pressure gradient near the ice edge.

1. INTRODUCTION

Koch and Wegener (1930) compared the mean monthly m.s.l. pressures at Borg (76° 41′N, 22° 25′W, 62 m) for the seven winter months of 1912-13 with the monthly m.s.l. pressure for Danmarkshavn (76° 46′N, 18° 46′W, 18 m) for the corresponding seven months for the years 1906-08, and showed that the Borg values were considerably higher than the Danmarkshavn values, the mean difference being 7·1 mb. They stated that Borg was 100 km nearer than Danmarkshavn to the centre of the glacial anticyclone, and that the 7·1 mb difference was very much larger than the difference (about 1 to 1·5 mb) one might expect over such a distance. Koch and Wegener concluded that the pressure gradient near the edge of the inland ice is especially great, but felt it was hardly possible that the whole of the 7·1 mb difference could be attributed to the pressure gradient, and concluded that in the winter 1912-13 the pressure in NE Greenland was a little higher than in 1906-08.

2. EXAMINATION OF RECENT DATA

(a) *Comparison of simultaneous pressure readings*

In 1952-54, pressure readings were taken at Britannia Sø (77° 09′N, 23° 36′W, altitude 229 m), approximately 135 km from Danmarkshavn in direction 285°, (Borg was approximately 105 km in direction 260°) simultaneously with those made at the same synoptic hours at Danmarkshavn.

The m.s.l. correction at Britannia Sø amounts to around 30 mb and there is some uncertainty as to what temperature to assume for the mean temperature of the 'air column between the station and the mean sea level.' The temperature adopted was the measured air temperature plus 1°C : an error of 1°C in this assumed temperature causes an error of 0·1 mb in the pressure correction. On account of inversions this correction may occasionally be about 0·5 mb in error, but it is unlikely in the mean that the error will exceed 0·3 mb.

I have extracted all the Danmarkshavn pressure, temperature and wind values which were plotted on the Atlantic charts at Prestwick for January 1953 and for which simultaneous values are available for Britannia Sø. These have been collected into two groups; (a) when the surface wind at Britannia Sø exceeded 15 kt and (b) when the surface wind at Britannia Sø was less than 5 kt. In addition, I excluded from (b) cases when the surface wind at Danmarkshavn exceeded 10 kt.

It was not until later that the altitude of Britannia Sø was determined and thus there are no pressure values for this station on the charts. On each occasion before computing the Britannia Sø m.s.l. pressure, I estimated the pressure difference p', between Britannia Sø and Danmarkshavn from the m.s.l. chart as drawn. This estimation is, of course, fairly rough as only few stations are available : nevertheless p' is a reasonably good estimate of the pressure difference which one would expect on the hypothesis that there is no unusual effect due to the edge of the ice sheet.

In group (a) (windy conditions) the mean difference, Britannia Sø – Danmarkshavn was 2·1 mb, the mean value of p' being 1·7 mb. In group (b) the mean difference was 1·1 mb, the mean value of p' being 0.

(b) *Comparison of monthly mean pressures*

To the mean station-level pressures for each month for Britannia Sø has been added the mean correction to m.s.l. given in the last column of Table 11 (b) of *Climatological Tables* (Hamilton and Rollitt 1957), and this corrected m.s.l. pressure for Britannia Sø has been compared with mean monthly pressures for Danmarkshavn. The difference Britannia Sø minus Danmarkshavn varied from a mean of 1·7 mb for the fine winter months of 1952-53, and a mean of 2·3 mb for the comparatively stormy winter of 1953-54 to about zero in the summer.

3. Conclusions

The two comparisons described above indicate that there is no unusually steep pressure gradient at the edge of the inland ice in NE Greenland. It seems that Koch and Wegener's result, quoted in the Introduction, was due to the fact that pressure was above the average in NE Greenland during the winter 1912-13. Such periods of high or low pressure occur in the Arctic, e.g., Scherhag (1957) has pointed out that in every month in 1953 there were pressure deficits in the Arctic region of 5 to 10 mb.

Acknowledgments

I wish to thank the Director, det Danske Meteorologiske Institut for supplying mean pressure values for Danmarkshavn, and the Director-General of the Meteorological Office for permission to publish this note.

References

Hamilton, R. A. and Rollitt, G. 1957 *Medd. Grønland*, **158**, No. 2.
Koch, J. P. and Wegener, A. 1930 *Ibid.*, **75**.
Scherhag, R. 1957 *J. Atmos. and Terr. Phys.*, Special Supp. 1957, p. 94.

551.543.1

The diurnal variation of pressure on the Greenland ice sheet

By R. A. HAMILTON
Meteorological Office, Prestwick

(Manuscript received 15 January 1959)

Summary

The curve of the diurnal pressure variation at ' Northice ' shows a marked minimum in summer and autumn at about 0430 GMT. Maxima occur at 1300 and 2100 separated by a subsidiary minimum at 1730 GMT. The range is 0·3 mb.

1. Introduction

Simpson (1918), discussing the 12-hourly pressure oscillation over the whole world, showed that this consisted of two components, the equatorial vibration which occurs at the same local time at all places on the same latitude, and the polar vibration which occurs at the same universal time at all stations at the same latitude. He derived equations in which the amplitudes and phases of the two vibrations were expressed as functions of the latitude and longitude. The amplitude of the equatorial amplitude decreases with latitude, and the polar vibration, though small with an amplitude of about 0·12 mb, predominates at high latitudes.

The quality and quantity of the data for the polar regions available to Simpson caused him to remark, ' When we come to the polar regions where the amplitude is almost infinitesimal we must expect large errors in the data with which we are working.' Wegener A. (1930) has pointed out that the diurnal pressure variation is so small that it can be masked by the effect of incomplete temperature compensation of the barograph if this is kept in an arctic hut which is subject to a large diurnal temperature variation. He expressed the opinion that all measurements of the daily pressure range in polar territory which are derived from aneroid barographs in heated huts must be treated with the greatest reserve. Holzapfel (1939) referring to the barograph at Umanak, remarks that ' it is perhaps of ethnographic interest as one can determine from it the time at which the eskimo observer, who as catechist had come under strong Danish cultural influence, started to heat his hut, but meteorologically it is valueless.'

Not only are most arctic huts subject to large diurnal temperature variations, but they are also subject to large temperature gradients between floor and ceiling. I have measured gradients of 10°C/m, and gradients exceeding 5°C/m must occur quite often. In these conditions it is doubtful if a barometer hanging near the wall of the hut is at a uniform temperature, and if the attached thermometer measures the temperature of the mercury column and scale. As an error of 1°C in the barometer temperature causes an error of nearly 0·2 mb in the pressure reading, temperature variations throughout the barometer will cause significant errors, and it is possible that the pressure measured by the barometer may be several tenths of a millibar in error.

It seems, then, that errors of up to 0·5 mb may occur in pressures measured by a barometer in an arctic hut, and as there is normally a diurnal routine of heating this may be reflected in the pressure readings. To study the diurnal pressure variation, therefore, it is advisable not to rely on any instrument in a heated hut. Wegener determined the shape of the diurnal pressure curve at Borg (76° 41′N, 22° 25′W, 62 m) for the first three months of 1913 from a barograph in a shaft in the firn (consolidated snow) beneath his winter house which was built on a glacier. In this shaft the temperature changed by only 0·1°C in a week. Georgi, Holzapfel and Kopp (1935) measured the pressure at ' Eismitte ' (70° 54′N, 40° 42′W, 3,000 m) by means of a mercury barometer and barograph in a cavern in the firn. This cavern was separated from the living-room (also a cavern in the firn) by a curtain, and the temperature variations were greater than at Borg. The barograph was almost perfectly compensated for temperature changes, however, and the barometer readings may be regarded as fairly reliable.

The accurate determination of a diurnal pressure curve requires observations over a long period, so that the effect of large pressure changes are compensated and eliminated. These firn cavern observations were made over relatively short periods, and it seems worth while, therefore, to supplement them by similar observations made at ' Northice ' (78° 04′N, 38° 27′W, 2341 m) on the Greenland ice sheet.

2. The barograph records

From June to December 1953 the barograph was housed in an alcove cut in the wall of the tunnel which led from the door of the hut, which was completely buried under the snow, to the

shaft leading up to the surface. The alcove was about 2 m below the surface. The barograph fitted snugly into the alcove which surrounded it on all sides, except for the glass front, with a separation of about 1 cm. The barograph was moved only once weekly when the chart was changed : time marks were made twice daily with a lever operated from outside. The range of the diurnal temperature range in the screen outside was from 7°C in the summer to negligibly small in the winter. During the summer the air in the tunnel was colder than that outside, so that the diurnal temperature range will have been small. No temperature measurements of the barograph were made, but it is reasonable to assume that the temperature variation was of the same order of magnitude as that measured by Wegener.

The barograph was similar to the standard M.O. open-scale barograph, but was specially made to work at a pressure of around 750 mb. The chart scale was 1 mb = 1·8 mm. Before being taken to Greenland the barograph passed the temperature compensation acceptance test, which requires the error to be less than 1·0 mb per 18°C, at normal pressure. No test was carried out at pressures around 750 mb. The trace was normally fine, and with the aid of a specially made scale and magnifying glass, could be estimated to 0·1 mb with an accuracy of $\pm$ 0·1 mb. The trace was rarely discontinuous at the time marks, suggesting that there was no appreciable friction.

The passage of fronts, ridges of high pressure, etc., caused pressure changes of the order of 20 mb and rapid changes in the rate-of-change of pressure. Over a short period of a month these would markedly affect the mean hourly pressure unless they were compensated by contrary values at another time. Accordingly, periods which could be seen from a glance at the barogram to be disturbed were excluded, and the analysis made of about a dozen 24-hr periods per month when the rate-of-change of pressure was almost linear. No rigid and objective criterion was used in picking out these undisturbed 24-hr periods, but a period would certainly be excluded if the barograph trace at any point differed by as much as 1 mb from the line joining the first and last points. The diurnal variation could not be detected by the eye and there could be no subjective choice of favourable periods.

3. Results

(*a*) '*Northice*'

The mean diurnal pressure curve for the summer-autumn period (June-November) is shown in Fig. 1. It is computed from the hourly readings on 79 undisturbed 24-hr periods. The main feature is the pronounced minimum around 0430 GMT (0200 LMT). The pressure is below the mean for only about 8 hr centred at the time of this minimum, and is above the mean for the rest of the day. The main maximum is around 1300 GMT with subsidiary minimum and maximum around 1730 and 2100 GMT respectively. The range is 0·30 mb.

It should be emphasized that except as a check on the range of the barograph, the barometer readings have not been used as a control on the barograph : the curves have been derived from the barograph trace alone.

At first sight, the mean barometer values (Hamilton and Rollitt 1957) do not appear to fit the diurnal pressure curves derived from the barograph, but the differences are so small that they can be accounted for by errors in the temperature correction. When the means of the barometer pressures for the same undisturbed days as were selected for analysis of the barograms are determined for the synoptic hours 06, 12 and 18 GMT for the six-month period, the values indicate a rise of 0·04 mb from 6 to 12 hr, and a fall of 0·09 mb from 12 to 18 hr. The diurnal pressure

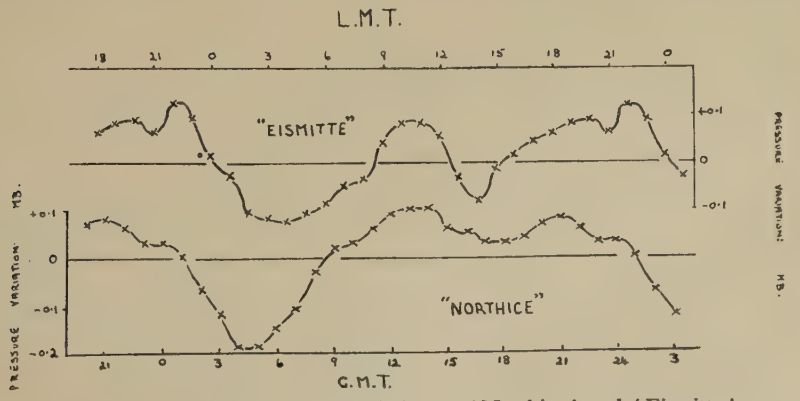


Figure 1. Diurnal pressure variation at 'Northice' and 'Eismitte.'

curve gives a rise of 0·25 mb from 6 to 12 hr and a fall of 0·06 mb from 12 to 18 hr. If, on an average, the mean temperature of the mercury column was 0·5°C higher than that of the attached thermometer at 6 hr, when room temperature was falling, and was 0·5°C lower at 12 and 18 hr when the room temperature was rising, the barometer would indicate a rise of 0·23 mb from 6 to 12 hr and a fall of 0·09 mb from 12 to 18 hr when correction for this systematic error is made. The diurnal range of temperature in the hut was about 8°C and small systematic differences of the order of temperature 0·5°C between the mercury column and the attached thermometer seem probable.

The diurnal range of temperature of the barograph was not known, but can hardly have been more than a few tenths of a degree even in the summer and must have been negligibly small in the winter. Even if an outside figure of 0·5°C is accepted, and the maximum error due to incomplete temperature compensation of 1 mb for 18°C is accepted, the resulting error in the diurnal pressure range is only 0·03 mb, i.e., one tenth of the measured range. This is well within the range of experimental error, and is so small that it can have little effect in determining the times of maximum and minimum.

(*b*) '*Eismitte*'

Holzapfel, Kopp and Wegener K. (1939) in working up the 'Eismitte' results, remarks that the curve of the daily pressure variation cannot be drawn as the non-periodic pressure variations in Greenland are large compared with those of the diurnal variation, and observations over one year would give only an inaccurate result. In his Table 12 the mean pressure values for each hour of the day are given for each month, but these values do not portray the diurnal variation as no correction has been made for gradual change of pressure during the month, e.g., the 01-hr and 24-hr values for December differ by as much as 0·7 mb, because there was a rise of 20 mb between 00 hr 1 December and 00 hr 1 January.

In order to obtain a diurnal pressure variation curve for 'Eismitte' I have plotted out all the hourly readings published by Georgi, and have selected 12 to 20 undisturbed 24-hr periods per month, as I did with the 'Northice' barograms. The mean curve for the period June-November is shown with the 'Northice' curve in Fig. 1. The 'Eismitte' curve shows the same characteristics as the 'Northice' curve, though the morning minimum is flatter, and about an hour later, and the afternoon minimum is more pronounced and occurs an hour earlier. The range is 0·23 mb compared with 0·30 mb for 'Northice.'

The 'Eismitte' hourly pressure values were obtained by interpolating from the barograph trace the values between those measured by the mercury barometer at 08, 14 and 21 hr LMT. The curve shows very marked minimum values at 14 and 21 hr, and one has the impression that the curve has been forced into this form by the barometer readings at 14 and 21 hr, and that a smoother curve with less pronounced afternoon minimum would have been obtained from the barograph trace alone. The 'Eismitte' barometer was not subject to as large temperature variations as the 'Northice' barometer, but nevertheless Georgi experienced trouble with temperature changes of the barometer because of the body heat of the observer. A systematic error of 0·5°C could account for these sharp minima.

4. Conclusions

The diurnal pressure variation on the inland ice in North Greenland has a range of about 0·3 mb. There is a pronounced minimum at around 0430 GMT (0200 LMT) and a subsidiary minimum around 1730 GMT. Maxima, which are approximately equal, occur about 1300 GMT and 2100 GMT.

Acknowledgment

I wish to thank the Director-General of the Meteorological Office for permission to publish this paper.

References

Georgi, J., Holzapfel, R. and Kopp, W. 1935 Wissenschaftliche Ergebnisse der Deutschen Grönland – Expedition Alfred Wegener 1929 and 1930/1931, Band IV, 1.

Hamilton, R. A. and Rollitt, G. 1957 Medd. om Grønland, **158**, 2, p. 53.

Holzapfel, R., Kopp, W. and Wegener, K. 1939 Wissenschaftliche Ergebnisse der Deutschen Grönland – Expedition Alfred Wegener 1929 and 1930/1931, Band IV, 2.

Simpson, G. C. 1918 *Quart. J. R. Met. Soc.*, **44**, p. 1.

Wegener, A. 1930 Medd. om Grønland, **75**, p. 473.

551.551.632.7

Some observations on flying locusts and atmospheric turbulence in eastern Africa

By R. C. RAINEY

(Read 17 December 1958. See *Q.J.*, **84**, p. 334)

Mr. F. H. Ludlam : Concerning the persistent coherence of the swarms, it does not appear to me to be unlikely that the atmospheric convection plays a part in preventing, rather than assisting, swarm dispersion. The swarms tend to be carried into belts of convergence in the surface winds, and judging by experience in England the convergence of the flow in such belts can be very considerable, commonly attaining values of about 10^{-4} sec^{-1}. Since the locusts cannot be carried up in the consequent vertical motion there must be a tendency for the swarms to shrink in horizontal area. Moreover, it is probable that because the sinking speed of the locusts relative to the air is comparable to the rising speeds of the more vigorous and local convective currents, these have the effect only of temporarily raising locust columns and do not have a marked effect in the horizontal diffusion of the insects.

Dr. R. S. Scorer : The question arises ' why do the locusts appear to be concentrated into sheets ? ' In any turbulent motion pairs of particles are, on the average, being separated because the motion diffuses airborne substance. Therefore at any point, at which any distortion can be seen to consist of an extension in any one direction, a contraction in a second, and a contraction or an extension in the third direction at right angles, in the case of turbulence on the average, the distortion in the third direction must be an extension. Thus compact volumes of fluid are drawn out, on the average, more often into sheets than into filaments. Townsend has shown experimentally that this is so. A second question is ' can a thermal concentrate falling bodies ? ' Particles falling through an isolated thermal are carried towards its vertical axis as they fall out of the bottom, and so particles which enter a thermal over the whole area of the top will fall out of the bottom as a narrow vertical column. This will apply to locusts gliding, and has been shown experimentally in the laboratory at Imperial College.

On the other hand, so long as the locusts remain in the thermal, which will be so long as its rate of rise is two or more times their sinking speed (ignoring their horizontal motion when at the edge of the thermal) they will tend to become widely distributed throughout the thermal.

Communicated : When interpreting the appearance of a swarm in terms of what a form of turbulent motion would do to a passive cloud of gliding locusts one must also ask whether the size of the group which naturally tends to fly all in the same direction will affect the appearance – if judged by the specular reflection from the wings. Could Dr. Rainey give any estimate of what this group size is, and is there a preferred natural period of time for which a locust will fly without change of direction except when emerging from the edge of a swarm ?

Dr. B. J. Mason : Dr. Rainey suggests that, in large dense swarms, the heat emitted by the locusts themselves may contribute appreciably to the buoyancy of the air; for locust concentrations of $0{\cdot}1/m^3$ he estimates a temperature rise of 0·1°C in 3 hours. It would seem that the effect of this heating would be more than offset by the loss of buoyancy due to the additional weight of the locusts amounting to, say, $0{\cdot}2$ g/m^3 and equivalent to a temperature drop of 0·05°C. Perhaps the continued cohesion of swarms over large distances may be explained by a locust's wishing to keep within visual range of its nearest neighbours while avoiding their wakes. Reasonable values for the visual range and the effect length of the wake would be 10 m and 25 cm respectively; these are compatible with concentrations ranging from $10^{-3}/m^3$ to $10^2/m^3$ quoted by Dr. Rainey.

Dr. A. G. Forsdyke : Dr. Rainey has suggested that the locust swarm, by its respiratory processes and interception of solar radiation, might impart heat to the ambient air sufficient to raise its temperature by 0·1°C in 3 hours, and thus contribute to the development of convective turbulence within the swarm. I suggest that this rise of temperature is so small in relation to that due to other causes, e.g., patchy heating of the ground on account of the variable nature and slope of the surface, or of cloud shadows, that it would not really affect the turbulence.

Prof. P. A. Sheppard (*President*) : The attempted explanation of the continued coherence of swarms, by Mr. Ludlam, in terms of convective ' bubble ' structure should not be pressed too far. The life of a convective bubble is quite short, a fraction of an hour presumably, whereas the swarm apparently retains its identity over many hours.

Dr. R. C. Rainey (*in reply, partly communicated*) : On any question of the extent to which a given meteorological process may account for a particular effect shown by swarms, it is, I think, important to consider the scale of the meteorological process, in time and space, relative to that of the effect shown by the locusts. This very useful discussion, for which my colleagues and I are most grateful, has in fact covered a wide range of such scales.

The smallest scale which has been considered is that of the detailed structure of high-flying swarms, involving elements of pattern characterized by lengths of order 0·1 to 1 km and by appreciable changes from minute to minute. Such dimensions are comparable with those of thermal bubbles; and the narrow vertical column in which falling particles can be expected to leave the base of such a bubble, as described by Dr. Scorer, may well correspond to the continuously-changing pillars of locusts which have been observed (*Q.J.*, **84**, p. 337). A possibly comparable effect, at a later stage of convective cloud development and on a correspondingly somewhat larger scale, is an association of individual swarms with individual scattered showers, noted, for example, in Somaliland in August 1957. Local convergence of surface wind-flow under such conditions may well have been at least comparable with Mr. Ludlam's figure (cf. e.g., Byers, H. R., 1949, *The thunderstorm*, Washington), though for attempting any rigorous assessment of such effects on locust distribution it would of course be necessary to consider not only instantaneous point values of convergence but also its effect as integrated over the time and area concerned.

Over this severely limited range of scales, I would therefore agree with Mr. Ludlam that atmospheric convection may oppose swarm dispersion. Swarm cohesion, on the other hand, is strikingly manifested on substantially larger scales, as Professor Sheppard points out, with individual swarms maintaining roughly constant diameters, of the order of 1-10 km, during displacements of hundreds of kilometers and over periods of weeks. I am not aware of any comparable effect shown by inert airborne material, nor of any purely meteorological process capable of accounting for such an effect, which I feel must therefore be attributed to the active behaviour of the locusts.

On a still larger scale are the movements and distribution of whole populations of swarms, over distances of thousands of kilometres during periods of months. Some of the corresponding patterns of tropical and sub-tropical surface wind-flow, on a continental scale, appear to show comparable persistence in time, and have exhibited a considerable measure of detailed agreement with the corresponding distributions of swarms (*Nature*, **168**, p. 1057).

On Mr. Ludlam's suggestion that swarms might tend to shrink in horizontal extent as a result of convective motion in the vertical, we have not so far been able to demonstrate any consistent effect of this kind. Thus, for example, aircraft traverses of the large Wajir swarms of 7 and 12 Jan. 1954, made during the afternoon with the locusts flying up to 1,200-1,700 m, gave swarm areas which were respectively 130 per cent and 60 per cent of the area of the same swarm similarly assessed early the same morning, while the locusts were still settled. On this point also I suspect that the behaviour of the locusts may be involved, and that they may be more tolerant of separation in the vertical than in the horizontal.

Dr. Scorer's description of a physical process by which turbulent dispersal could draw out compact volumes of fluid into thin sheets may clearly be highly relevant to the formation of sheets of locusts, though here again it would be necessary to know more about the scales on which the process which he describes could be expected to operate in the atmosphere, as well as more about the active reactions of flying locusts to each other which we postulate, before the relative contributions of these physical and biological processes could be assessed.

Concerning effects of specular reflection, it has been noted that, in relatively low-flying swarms viewed from above, the different groups of flying locusts, each probably of relatively uniform orientation, could on occasion be distinguished by differences in appearance of the light reflected off them; and, when large high-flying swarms were viewed, from close to, by specular reflection, only a small proportion of the swarm was visible at any one time (Fig. 2). Individual groups of uniformly-orientated locusts, as recorded photographically in passing swarms, have ranged from less than 0·1 to 0·5 km in length. We have no evidence of any preferred length of flight between changes of orientation, apart from those at the edge of the swarm.

In reply to Dr. Mason, I agree that any contribution to the buoyancy of the air provided by heat from the flying locusts must be set against the loss of buoyancy due to their weight (cf. p. 349). From my estimate of 0·1°C/3 hr for the former and his figure of 0·05°C for the latter, the additional buoyancy due to the heat imparted to the air by flying locusts could be expected – in a hypothetical completely closed system – to equal the loss of buoyancy due to the weight of the locusts after about 1½ hr; and this latter figure should be about the same for any constant concentration (volume density) of flying locusts. In practice, I should expect the effects of heating by the locusts to depend

on whether or not the air with the locusts in it approximated to a closed system. Thus with a small stratiform swarm, travelling at less than the speed of the wind, the heat from the locusts may well be lost so rapidly, across the perimeter and the upper surface of the swarm, that its effects on buoyancy would be more than offset by the weight of the flying locusts, as Dr. Mason suggests. On the other hand, in the centre of a large cumuliform swarm, over 10 km in diameter, and travelling at the speed of the wind for 8 hr or more at a stretch, conditions should be much more favourable for the accumulation of sufficient heat from the locusts to exceed the loss of buoyancy due to their weight, perhaps to the extent of a net excess of a few tenths of a degree by the late afternoon. It is tempting to suggest that such an effect might have contributed to a subjective but repeated observation that the apparent proportion of high-flying locusts reached a marked maximum in the late afternoon.

On Dr. Forsdyke's query as to whether so small a difference of temperature could have any significant effect on the stability of the air, I agree that the surface of the ground traversed by these swarms shows much larger differences of temperature, within distances of metres or less. The small temperature difference which the locusts could be expected to produce, on the other hand, might be established within a mass of air covering many km^2, of the order of km deep, and might be maintained for a matter of hours. Since the ascent of an ordinary thermal bubble, a few hundred metres in diameter, is due to a temperature excess of the order of a degree, while buoyancy forces and the corresponding drag involve respectively cubes and squares of linear dimensions, I suggest that a temperature excess considerably less than a degree might still be significant in relation to the substantially larger volumes concerned with such swarms.

Finally, Dr. Mason's suggestion that the continued cohesion of travelling swarms, and the range of concentrations of flying locusts observed in them, might be interpreted in terms of behaviour reactions, operating to keep each locust within visual range of its neighbours while at the same time avoiding their wakes, is a most attractive hypothesis. It is, as he points out, in order-of-magnitude agreement with the quantitative data so far available, though I should perhaps emphasize the limitations of the latter.

551.524.4 : 595

The vertical distribution of aphids in the air and the temperature lapse rate

By C. G. JOHNSON

(Read 17 December 1958. Presented by H. L. Penman and questions answered by L. R. Taylor. See *Q.J.*, **83**, p. 194)

Dr. H. L. Penman added the following analytical notes in the course of his presentation :

It is known that the initial vigour of flight of aphids is not long sustained, and thereafter it may be safe to treat them as inert particles whose motion is a compound of turbulent transport upward, and a falling-out downward. The net upward flux will be given by

$$F = -K\,\partial\rho/\partial z + w\rho$$

where K is the coefficient of eddy diffusion $= k\,u_*\,z$

ρ is the insect population per unit volume

w is the upward velocity of the aphids relative to the air,

and k and u_* have their conventional meanings. As an assumption, to make progress, set $w\rho = \lambda K\,\partial\rho/\partial z$, where λ is independent of z and ρ : from which $F = -K(1-\lambda)\,\partial\rho/\partial z$, and $K(1-\lambda)$ is an effective transport constant. If ordinary diffusion theory holds over the range of Johnson's observations then it is possible to derive

$$\ln\rho = b\ln z + \text{constant}$$

where

$$b = w/\lambda k\,u_*,$$

i.e., plotting log ρ against log z should give a straight line of slope b, the result demonstrated in Fig. 4 of the paper. From $bu_* = w/\lambda k$, if w and λ are approximately constant, we can infer – qualitatively – the results set out in Fig. 1 and 2 of the paper.

The remainder of Johnson's data offers the possibility of estimating w and λ. Fig. 5 of the paper shows a marked contrast between the early morning and mid-day density curves. For the latter, Johnson suggests that there is a buoyancy transfer, the aphids caught at 500 ft being those that might have been caught at 10 ft a few minutes earlier. For the first peak of the day there is the near certainty that the transport is solely due to turbulence, and though the wave is a transient we might represent it as being transmitted according to

$$\rho = \rho_0 + A\, e^{-(a+ib)z}\, e^{2\pi it/T}$$

where a and b are related to the diffusivity, and T is the period of the wave.

Applying an analysis (McCulloch and Penman 1956) for diffusivity (K') varying with z, and making the guess that keeps the algebra as simple as possible by postulating $a = \alpha z^{-\frac{1}{2}}$, $b = \beta z^{-\frac{1}{2}}$, then the following emerge :

$$K' = \frac{2\pi}{T}\frac{4\alpha z}{\beta(\alpha^2+\beta^2)} = K(1-\lambda)$$

$$\alpha = \ln A_1/A_2 \Big/ (z_2^{\frac{1}{2}} - z_1^{\frac{1}{2}})$$

$$\beta = \frac{2\pi(t_2 - t_1)}{T} \Big/ (z_2^{\frac{1}{2}} - z_1^{\frac{1}{2}})$$

where A_1 and A_2 are the amplitudes at z_1 and z_2, and $(t_2 - t_1)$ is the interval between the occurrences of maxima at z_2 and z_1.

From Fig. 5 of the paper roughly, $A_1/A_2 = 200/30$, $t_2 - t_1 = 2{\cdot}3$ hr, $z_2^{\frac{1}{2}} - z_1^{\frac{1}{2}} = 19{\cdot}2$ ft$^{\frac{1}{2}}$.

Hence $\alpha \simeq 0{\cdot}10$

$\beta \simeq 0{\cdot}76/T$ with T in hours.

If the build-up is regarded as a half period, $T \simeq 4$ hr : if as a quarter period, $T \simeq 8$ hr. The corresponding estimates of K' are

$$K' = 0{\cdot}017 \text{ ft}^2 \text{ sec}^{-1}, \quad T = 4$$

$$K' = 0{\cdot}04 \text{ ft}^2 \text{ sec}^{-1}, \quad T = 8$$

From the wind data at 1,000 and 135 ft, given by Johnson, $K = 0{\cdot}12\, z$ ft^2 sec^{-1}, and hence $\lambda = 1 - K'/K = 0{\cdot}67$ or $0{\cdot}86$. It is safe to assume that b is close to $-1{\cdot}0$, and hence

$$w = -0{\cdot}08 \text{ or } -0{\cdot}10 \text{ ft sec}^{-1}$$

$$= 2{\cdot}4 \text{ or } 3{\cdot}0 \text{ cm sec}^{-1}, \text{ downward.}$$

This seems to be an acceptable order of magnitude for an aphid with spread wings : if so, then λ also may be about the right size, i.e., close to the value of unity which would represent equilibrium between upward turbulent diffusion and downward fall-out. Hence Johnson's reference to his log/log plots as being for quasi-stationary states seems to be fair comment.

Reference

McCulloch, J. S. G. and Penman, H. L. 1956 'Heat flow in the soil,' *Rept. 6th Int. Congr. Soil Sci.*, Paris, Vol. B, p. 275.

Mr. W. G. Harper : On many occasions aphids must be carried aloft in thermals. Does Mr. Taylor think that, as in the case of Dr. Rainey's locusts, aphids will then be found at all heights up to inversion levels ? It seems possible that near inversions, the aphids might achieve quite long drift flights if their urge to fly persists, and might even become concentrated near inversion levels. Is there any evidence for such concentrations, and is it possible that the rapid fall in numbers at around 1400 hr in Dr. Johnson's Fig. 3 was due to the aphids being convected most rapidly upwards at that time, and that there could even have been a maximum in numbers *above* 1,000 ft just when the minimum was occurring near the ground ?

Dr. R. C. Rainey (*partly communicated*) : On 7 Aug. 1924 spruce aphids (*Dilachnus piceae* Pz.) arrived in very large numbers over an extensive area of the ice-cap of North-East Land, Spitzbergen. The nearest area of extensive spruce, capable of providing an adquate source for the aphids, was the Kola peninsula, over 800 mi away to the SSE; and there was a spell of strong southerly and south-easterly winds over this area during 5-7 Aug. (Elton 1925, *Trans. Roy. Ent. Soc.*, London, pp. 289-299). I should like to ask whether Dr. Johnson has any comparable trapping records, for species of aphid with a host-plant range similarly indicating a distant origin, which might conversely provide evidence of the trajectory of the air concerned.

Mrs. K. J. Richardson : If more than one species is concerned in a migration, identification becomes very important, because their behaviour may differ a lot.

I remember watching migrating butterflies in Ceylon. On one occasion, three separate migrations of three species were going on in different directions at the same time and in the same place. There can be no question that their very different behaviour was due to specific differences and not to climatic conditions. (See *Proc. Roy. Ent. Soc.*, London, **20**, 1945, pp. 114-115).

Mr. J. G. Skellam (*partly communicated*) : In the simple model which Dr. Penman has outlined, the parameters are independent of time. Such a system would provide for an advancing ' wave ' in the early part of the day but not for a return ' wave ' or general collapse towards the end of the day. If, however, the natural tendency of the aphids to descend under gravity were to increase before the end of the day as a result of some change in the behaviour such as the cessation of flight, we could well expect to obtain results similar to those described. The distribution of density in the lower traps would be bimodal, this bimodality being less marked with height and untimately disappearing. (A later speaker drew attention to the bimodality of the integrated curve, but this opens another question – that of the validity of the integrated values : whether in computing the integral the distribution has suffered truncation, or whether a method of extrapolation has been employed which is free from objection).

Mr. J. S. Sawyer : The double peak in the diurnal curve of aphid density recalls the similar feature which is commonly found in curves of the daily variation of atmospheric humidity in the lowest few hundred feet. The subsidiary minimum in humidity in the late morning is generally attributed to the effect of the increasing upward transport by turbulence outweighing evaporation from the ground, and the second (afternoon) maximum to the result of the reduced convection combined with evaporation continuing at a high level. Although there may be a complicating diurnal rhythm in the number of aphids which take to flight, it would be worth comparing the distribution and variation of aphid density with that of specific humidity on the same occasion; close parallelism would suggest that atmospheric effects dominate the diurnal variation of aphid density.

D. K. Mellanby : For an entomologist, it is interesting to compare the two species of insect which have been discussed today. The locust is one of the largest insects, a powerful flier and one which appears to ' know where it is going.' The aphid is a very small, feeble insect, a poor flier, and one that would appear to have little control over its movements in the air. Yet we see that when the effects of meteorological factors are studied, there is much more that is common in their effects on these insects than might previously have been expected from the studies that have been made of the morphology, and even of the physiology, of the locust and the aphid.

Prof. P. A. Sheppard (*President*) : The diurnal variation of aphid concentrations must presumably become a single-maximum curve at higher elevations, if the variation is to be accounted for in meteorological terms. Is there evidence of this ? If not, then a marked diurnal variation of ground-source strength of aphids would appear to be demanded by the curves shown.

Mr. L. R. Taylor (*in reply*) : In answer to Dr. Rainey I should explain that not all the trap catches from high levels have yet been identified. None of the insects so far dealt with could be said, with confidence, to have come from any great distance. It should, however, be remembered that knowledge of aphid distribution and food plants is very imperfect. This, combined with the very varied vegetation of this region, makes the chances of identification of the origin of trapped aphids very much less than it appears from Elton's very fortunate encounter.

Mrs. Richardson's experience with migrating butterflies is an example of the controlled flight of relatively large insects with flight speeds in excess of the air speed. In most circumstances aphids are flying in a medium which moves faster than they can fly, and specific differences in behaviour have little effect on their distribution. It has also been shown that the periodicity of flight of many species of migrant aphids is very similar.

It is perhaps best to answer Mr. Skellam, Mr. Harper and Mr. Sawyer together since their questions are inter-related. In the material at present available there is no evidence of the bimodality disappearing with height; in the example shown in Fig. 3 of Dr. Johnson's paper the second

maximum does not reach even to 1,000 ft. Also a close examination of Fig. 4 will show that the minimum at 15.00 is brought about, not by attenuation of low-level densities as would be expected if increasing rate of upward transport outpaced supply, but by loss of insects from the upper layers and a consequent *increase* in the gradient of the profile. Again the afternoon maximum is built up by an increase in density at *low* levels whilst maintaining this high gradient; this could surely result only from a replenishment of insects from the ground ?

If this is so, then the question raised by Mr. Skellam about the validity of the extrapolation used to produce an integrated total insect content of the air is of less importance. In any case, Dr. Johnson's extrapolations were extended up to the height at which the density equals 1 per 10^6 cu ft of air : I have since repeated the operation using extrapolations up to a standard height of 10,000 ft, with precisely the same result. It seems likely that these two extremes encompass the truth, unless there are accumulations of insects in isolated pockets, perhaps in inversions, and these could hardly affect the results.

This brings me to Mr. Harper's point about accumulation. As pointed out above, such accumulations above 1,000 ft could not be detected by our methods, and the only evidence we have is biological. It is well established that the behaviour of migratory aphids changes radically after a very short period of flight and the aphids then lose lift and would be expected to settle out unless some external factor caused them to remain in the air. This argument may not apply to some other insects and we have at least one example of a sustained flight at a particular level. The fruit fly, *Drosophila*, which is known to fly normally only at twilight, was found flying until after midnight at 56 ft when the air below was already cleared of these insects.

As Dr. Mellanby reminds us, the similarity between the way in which both aphids and locusts make use of the movement of air to assist in dispersal is at first sight surprising, in view of their morphological differences. But if they are both viewed in the light of the known dispersal of plant spores at the one extreme and the use made of air movements by the larger gliding birds at the other, they are then seen to form part of a much longer series.

In reply to Prof. Sheppard, I would say that, as mentioned earlier, there is no reason to believe that the single maximum curve exists at higher levels. On the other hand, the bimodal curve of production of new aphids ready for their first, and possibly only, flight is very well established. Also the behaviour of aphids, which leads new adults to fly – but for a very short time only – is also well investigated. It therefore seems more likely that the aphids return to earth during the mid-day minimum than that they move to higher levels.

Dr. H. L. Penman (*in reply*) : I have nothing to add to Mr. Taylor's replies, other than an appeal. This useful discussion will have its greatest value if it provokes someone into making a more expert analysis of the meteorological physics of the problem on a less flimsy basis than that used by me. Dr. Johnson has put into this paper only a small fraction of the data he possesses, and all can be made available to anyone who wishes to match theory and observation.

551.506.5 (988) : 551.589.1

The meteorology of North Greenland during the midsummer period

and

551.506.5 (988) : 551.589.1

The meteorology of North Greenland during the midwinter period

By R. A. HAMILTON

(Read 21 January 1959. See *Q.J.*, **84**, pp. 142 and 355)

Prof. R. G. Fleagle : In the abstract of Mr. Hamilton's paper there is a reference to katabatic winds flowing from the higher parts of the Greenland ice sheet. I should like to ask whether his observations were made with detail sufficient to describe the changes in katabatic wind speed with time and the relation between slope of the surface and speed of the wind. Specifically, was there any evidence of periodicity in the early stages of the katabatic wind ?

Mr. G. J. Jefferson : I note that Mr. Hamilton has used 700 mb contour charts in the same way as surface charts are used elsewhere for analysis by drawing on fronts, etc. I do not know whether he actually drew the charts on the spot or during the working up of the data afterwards, but was any attempt made at a three-dimensional understanding of the air masses over Greenland by the use of frontal contour charts ? It would appear that a method of this sort might be of special value in such an area.

In the course of some work which I have been doing I have examined daily synoptic charts for the surface and for the 500 mb level for a period of some years and it appears that depressions cross North Greenland quite frequently. What usually appears to happen is that quite a deep depression shown on the surface chart off the northern part of the west coast of Greenland accompanied by a depression or sometimes only a trough at 500 mb fills up quite rapidly as the upper depression or trough moves eastwards across the ice cap. When this reaches the east coast a further surface depression, often quite deep, develops and moves away eastward. It is as though the depression has moved across Greenland in what one might call an ' inhibited ' form. Can Mr. Hamilton say whether, in his experience, this is of common occurrence ?

Prof. G. Manley : I should like to ask for information on a matter of old-fashioned climatology. I think it was in 1877 that one of the most astonishing meteorological events on record occurred, namely, a February maximum of 61°F at Upernivik which lies about 73°N on the west coast. Obviously air must have been transported right across the ice cap from somewhere south of Iceland and any inversion layer normally existing must have been totally removed. Does Mr. Hamilton know of any occasion when unusually high winter temperatures have occurred, say, round Independence Fjord or elsewhere in the far north-east. Presumably they would derive from a surge of air originating in the open water of Davis Strait and would be reflected in an unusual rise of temperature at ' Northice,' associated with a depression travelling across the Arctic Basin from the north of Ellesmere Land; a route for which there is some evidence.

Mr. J. S. Sawyer : Can Mr. Hamilton tell us whether the ice-cap observations support the idea that wind streams directed towards Greenland frequently or occasionally surmount the ice-cap or whether they are partially or totally deflected around its flanks ? A comparison between temperatures over the ice-cap and radiosonde observations on the coast should give a clue.

Mr. M. K. Miles : From maps of thickness (1,000-500 mb) I get the impression that deep tropospheric cold outbreaks come southwards on either side of Greenland but not across it. It is very uncommon, I think, to find Angmakssalik reporting an appreciably lower thickness than places to east and west of it.

Would Mr. Hamilton care to comment on this in the light of his study of the meteorology of Greenland ?

Mr. L. C. W. Bonacina : I have often wondered whether if the North Polar Basin were land at sea level, instead of being filled with sea-ice as at present, it would maintain an ice sheet. If there was no ice to begin with I doubt whether the snowfall in present climatic conditions would be sufficient to initiate an ice sheet. An incipient ice sheet might not survive the summer sun circling round the sky at an average altitude of about 30 degrees between May and August. A critical balance of factors would be involved about which I invite comment.

Mr. H. H. Lamb : I consider that the type of Arctic expedition most closely related to this country's economic, as well as meteorological, interests would be a maritime expedition operating in the Norwegian and Barents Seas. The aim should be to learn more about the meteorological processes over those waters, e.g., modification of airmasses arriving from Greenland and from the polar pack-ice, the relative importance of these respective areas as sources of vigorous cold airstreams, etc., with a view to improving the weather protection service for the Arctic fisheries. (The German fishery protection vessels already mount a routine forecasting service in northern waters, but more remains to be done on the investigational side).

Referring to a previous speaker's remarks on the role of Greenland in the upper-air circulation, I believe the general impression gained from 500 mb topography, 1,000-500 mb thickness and surface weather studies is that tongues of deep cold air advance either west or east of Greenland – not over or from Greenland. That is to say, Greenland appears to act more as a barrier than as a cold source. In this respect Antarctica differs, being quite clearly an important cold source from which deep cold airmasses emerge over the ocean.

Dr. D. J. Schove : I have been very interested in these two papers and I am wondering to what extent precipitation on the one hand and snow accumulation on the other can be linked with the pressure anomaly map. I am intrigued with the measurement of annual layers, which go back centuries, and wonder whether these will throw any light on the pressure patterns of individual years.

Mr. R. A. Hamilton : In reply to Prof. Fleagle, I would say that it is very desirable that measurements of the speed and depth of the katabatic wind should be made at open sites at several stations on the inland ice, particularly near its periphery. As written in my paper (p. 364) the NW-ly wind at Britannia Sø is not a true katabatic wind. There was no clear evidence of periodicity in its early stages. At ' Northice ' it blew almost continuously, its speed being related to the pressure distribution.

I have not tried to use frontal contour charts, which may well be useful. The work of analysis is rather handicapped by the frequent lack of humidity data on the radiosonde reports. Our experience is that it is a common occurrence for depressions to cross Greenland, and at the 700 mb level there is little if any deflection of the wind streams.

Prof. Manley has probably quoted an extreme case of the high temperatures observed in winter on the coast in an air mass which has crossed the inland ice, such as I referred to on p. 315.

I think Mr. Lamb has answered Mr. Miles's question, for which I am grateful as I could not. In my introductory remarks I made it clear that I have tried hard to get the meteorological results of the expedition worked up in reasonable time, and I have therefore concentrated on doing this. I have not studied the 500 mb charts of the Greenland area, and make no claim to have made an exhaustive study of the meteorology of Greenland.

Mr. Bonacina's question brings us into the pleasant realms of fantasy in which one cannot be proved wrong ! My opinion is that an ice-sheet would not develop if the Arctic Basin were low-lying land. Most of the winter's deposit of snow on the polar pack melts during the summer, and snow disappears quite early in the summer from most of the low-lying lands around the Arctic Ocean. In these lands summer temperatures are higher than in the polar basin where the water prevents the air temperature from exceeding 0°C appreciably.

I do not know enough about it to attempt to reply to Dr. Schove, but just throw in this warning : do not get confused with precipitation as measured by the meteorologist, and accumulation as measured by the glaciologist, for they can be markedly different. As accumulation depends on the wind speed, one must be careful, in linking accumulation in past years with pressure anomalies, not to jump to conclusions about precipitation in these years.

Proceedings at meetings of the Society

Evening Discourse, 7 January 1959

The second of the series of Evening Discourses took place in the Society's Rooms : Dr. R. C. Sutcliffe, F.R.S., spoke on ' The future of weather forecasting.'

A full account of this meeting appeared in March 1959 issue of *Weather*.

Ordinary Meeting, 21 January 1959

The following candidates were declared to have been elected that day by the Council :

FELLOWS

COX, Frederick Royou, 67 Scarle Road, Wembley, Middx.
GROOCOCK, John Anthony, 117 Derbyshire Lane, Sheffield, 8, Yorkshire.
MILFORD, James Richard, B.A., Peterhouse, Marandellas, S. Rhodesia.
PHILIP, John Robert, B.C.E., C.S.I.R.O., Division of Plane Industry, P.O. Box 109, City, Canberra, Australia.

STUDENTS

COATES, Richard Martin, 78 Woodberry Avenue, North Harrow, Middx.
EL-MASRI, M. Hassan, Meteorology Dept., Imperial College, Exhibition Road, London, S.W.7.
HUTCHINSON, Anthony, ' Mundworth,' 230 Leigham Court Road, Streatham, London, S.W.16.

The following papers were read and discussed :

1. ' Some measurements on undercutting cold air ' by F. A. Berson (*Q.J.*, **84**, p. 1) presented by Mr. M. K. Miles.
2. ' The meteorology of North Greenland during the midsummer period ' by R. A. Hamilton (*Q.J.*, **84**, p. 142).
3. ' The meteorology of North Greenland during the midwinter period ' by R. A. Hamilton (*Q.J.*, **84**, p.355).

Special General Meeting, 18 February 1959

Professor P. A. Sheppard, President, in the Chair, and about 40 Fellows present.

The Business of this Meeting which had been postponed from a Special General Meeting, called on 17 December 1958, was for the purpose of amending certain of the Society's By-laws; firstly to include a new grade of Corporate Member, whereby business concerns and other bodies might be enabled to support the Society's activities; and secondly to provide for a reduced subscription in the case of Members of long standing who had retired from paid employment.

To give effect to the desired changes the following amendments to the By-laws were proposed :

(A number in brackets is in all cases the new number of a By-law, resulting from the deletion of old paragraphs or from the insertion of new ones).

By-law 5
After ' Fellows ' insert ' Corporate Members.' Delete ' Associates.'

By-law 6
Delete.

By-law 7 (6)
After ' Fellows ' insert ' Corporate Members.' Delete ' Associates.'

By-law 9 (8)
Replace by ' Each Corporate Member shall nominate one person who shall act for the Corporate Member in all dealings and correspondence with the Society.'

By-law 10 (9)
Line 1 and line 6. Replace ' Associate ' by ' Corporate Member.' Delete ' and the nominees of Instititions, Companies or Corporate Bodies.'

By-law 11 (10)
Line 3. Replace ' Associates ' by ' and.' Delete ' and nominees of Corporate Bodies.'
Insert new By-law 11
' The Society may admit such Corporate Bodies as may be qualified and elected in that behalf as Corporate Members, and Form A shall be signed on behalf of such Corporate Member by the person nominated under By-law 8.'

By-law 16 Re-word as follows :

Corporate Members

' Any school, college or university, or any department of a school, college or university, any Local or County Council, Research Institute, Company or other Corporate Body having an interest in meteorology shall be eligible as a candidate for election to the class of Corporate Member.'

By-law 21
Delete ' or Associateship.'

By-law 35
Delete ' Associates £1. 1s.'

By-law 36
Replace ' Associates £3. 0s.' by ' Corporate Members £21.' Delete ' except an institution or corporate body.'

Insert new By-law 37
' Fellows residing outside the British Isles may if they so desire pay a reduced annual subscription of 75 per cent of the full rate.'

Insert new By-law 38
' Any Fellow or Foreign Member who is 65 years of age or more and has been a Member for 30 years or more is entitled, on written application to the Council, to pay an annual subscription of 50 per cent of the full rate.'

(The numbers of all By-laws, commencing with the present By-law 37, to be increased by 2).

By-law 49 (51) line 2 and By-law 51 (53) lines 2-3
Delete ' entitled to vote.'

By-law 68 (70)
Replace ' member of any class ' by ' Honorary Member, Fellow, Foreign Member or Student.' Add at the end of By-law ' Up to five representatives of a Corporate Member may attend any Ordinary Meeting. Permission for the attendance of more than five representatives must be obtained from a Secretary and will be at his discretion.'

By-law 73 (75)
Delete ' and entitled to vote at such Meeting.'

By-law 74 (76)
Replace by ' Only a Fellow shall be appointed as proxy to vote at any meeting.'

The President explained the policy underlying the proposed changes. A short discussion followed and the President then proposed the Resolution which was carried unanimously.

Ordinary Meeting, 18 February 1959

The following candidates were declared to have been elected that day by the Council :

FELLOWS

ALLINGTON, Kathleen R., B.Sc., c/o Department of Geography, McGill University, Montreal, Que., Canada (*transfer from Student Membership*).
ASTON, Vivian, 166 Margam Road, Port Talbot, S. Wales (*transfer from Student Membership*).
GRIFFIN, Trevor Louis Charles, B.A., 107 King Edward's Road, Swansea, Glamorgan, S. Wales.
HARIHAR, Krishna, B.Sc., Vimal Niketan, 10th Road, Khar, Bombay 21, India.
PEUGNIEZ, Patrick Julian, B.Sc., Hillrise, Allendale Road, Hexham, Northumberland (*transfer from Student Membership*).
RAO, Yadavalli Paramasiva, M.Sc., Meteorologist-in-Charge, Bombay Airport, Bombay 29, India.
SHAM, Patrick Pak, B.Sc., 1D Belcher Gardens, Hong Kong.

FOREIGN MEMBERS

ASHBURN, Edward Victor, M.S., Naval Ordnance Test Station, Code 5021, China Lake, California, U.S.A.
BARRIENTOS, Celso S., B.S., Bayanan, Bacoor, Cavite, Philippines.
LUDLAM, David M., Ph.D., 285 Riverside Drive, Princeton, New Jersey, U.S.A.
MAKANAS, Alvier D., B.S., Weather Bureau, Manila, Philippines.
SANGSTER, Wayne Eugene, M.S., Department of Meteorology, University of Chicago, Chicago 37, Illinois, U.S.A.

STUDENT MEMBER

NEWBOLD, David Hesketh, 16 Drayton Gardens, West Ealing, London, W.13.

The following papers were read and discussed :

1. ' The temperature, humidity and cloud near fronts in the middle and upper troposphere ' by J. S. Sawyer (*Q.J.*, **84**, p. 375).
2. ' Long waves, lee waves and gravity waves ' by F. K. Ball (*Q.J.*, **85**, p. 24), presented by Mr. G. A. Corby.
3. ' Ice-nucleating properties of some natural mineral dusts ' by B. J. Mason and J. Maybank (*Q.J.*, **84**, p. 235).

Evening Discourse, 4 March 1959

The third of the series of Evening Discourses took place in the Society's Rooms : Mr. H. H. Lamb spoke on ' Our changing climate, past and present.'

A full account of this meeting will appear in *Weather*.

Polar atmosphere symposium, Pt. 1, Meteorology section. Edited by R. C. Sutcliffe. London (Pergamon Press, for, and on behalf of, A.G.A.R.D.), 1958. Pp. xi, 341; 193 Figs.; 92 Tables. 70s.

This handsomely produced book, together with its companion volume (*Pt. 2, Ionospheric section*, 60s.), is the record of a meeting. It presents the papers read at a symposium held in Oslo from 2-8 July 1956, including the discussions at the end of each session.

Probably few readers will have previously heard of A.G.A.R.D., the Advisory Group for Aeronautical Research and Development of the North Atlantic Treaty Organization, whose objectives are intended to include activities serving the development of peaceful and friendly international relations (Article 2 of the Treaty). This particular symposium was suggested and planned by Norway because of that country's special concern with the problems of the polar atmosphere. So many leading figures in meteorology in the western nations took part that the discussions have an unusual interest for workers in polar meteorology.

Inevitably, the proceedings of a series of meetings of this nature make a rather curious book. The relative lengths of different papers do not correspond to the importance of their respective themes, but most are full of ' meat '; and there is much that is new and not yet published in any text-book. There is some duplication, for instance, the breakdown of the circumpolar circulation at the 50-mb level (*circa* 20 km in the ozone layer) in February 1952, the occasion of the ' Berlin warming,' is analysed by chart sequences in two papers in different parts of the book. This, and other topics, are returned to in several different discussions.

The serious student will find it worth while to compile an index for himself. The subjects treated range from the optics of the Arctic ' white-out,' to a comparison of the large-scale circulations around the northern and southern polar regions, and recommendations regarding which of the published daily weather analyses come nearest to being reliable over the Arctic; from intensive study of the day-by-day development of inversion structure over an Arctic ice island to mathematical representations of the flow patterns and pressure fields over the whole of the polar region. Sverdrup made some interesting points about the different albedos of Arctic and Antarctic pack-ice and its snow-cover. The reader will also learn that the American, as well as the Russian sector of the Arctic, now has its organization for surveying and forecasting the sea ice. Examples of its work are presented.

Very few misprints of any importance were detected, but quotation of the year 1870 (on p.27) for the beginning of the recent climatic warming in Europe and decrease of ice at the coasts of Iceland seems to be misleading (possibly 1890 was intended). The reviewer would also have liked to see formally complete legends for several of the figures.

There was a fairly general feeling that the Canadian three-front model could lead to too much rigidity in routine analysis unless provision were constantly made for fronts continually undergoing dissolution or sharpening. Nevertheless, fronts are real, and there should be no reaction towards abandoning the concept. Several interesting vertical cross-sections in the Alaskan sector were shown by Reed, who seemed however to minimize the evidence on his diagrams for fronts, especially weak fronts with occlusion structure, which are common in the Arctic and Antarctic. Russian cross-section analysis (not illustrated at the meeting) supports the idea of frequent weak, and very occasional strong high-reaching, fronts in the Arctic.

Other points which the reviewer found specially interesting illustrate the diversity of the book's content. Difficulties of observing include conditions which show up the defects of the Stevenson screen, which may indicate temperatures in summer up to 4°C higher than the Assmann psychrometer with its controlled ventilation. This is partly because of the intensity of the reflected radiation from the snow below. On a few of the best days in summer, the radiation actually reaching the polar ice exceeds that available in low latitudes. The extreme transparency of the polar atmosphere contributes to this. On a small percentage of occasions there is no inversion in the lower troposphere. Overcast skies in early summer are less opaque than later on, when moisture content has increased. Associations have been noted between occasions when the cold air of the northern polar cap becomes displaced over North America and the subsequent conditions both in Europe (blocking) and over the polar region (warming). Traces of diurnal variations of various elements in the Arctic and Antarctic, sometimes with the mean maxima between midnight and 4 a.m. appear, as yet, quite unexplained.

This is a book which should be in the hands of everyone concerned with polar meteorology. Neither expedition planners, scientists in charge of observations at the bases, nor research workers touching upon this field at home, can afford to be ignorant of its contents. Unavoidably the price is no doubt high for individuals to rise to, but it may be hoped that all the appropriate libraries will have it and that copies will find their way to the bases of future expeditions.

H.H.L.

The upper atmosphere. By H. S. W. Massey and R. F. L. Boyd. London (Hutchinson and Co.), 1958. Pp. xii, 333; 135 Figs.; 6 Tables; 28 Plates. 63s.

Although the main interest in investigations of the upper atmosphere, the region from say 20 km altitude to outer space, is at present centred on its gross physical characteristics it is becoming evident that the time is not far ahead when it will be possible for meteorologists to study in some detail the synoptic and dynamic conditions of this region. The publication of this authoritative, up-to-date and easily readable account of the present state of knowledge and the researches now in progress is therefore very welcome and provides a background to which meteorological ideas can be related. Its value as an introduction to other subjects such as radio transmission, solar-terrestrial relationships, the earth's magnetic field, and the atomic and molecular

physics of upper atmospheric processes is also obvious. Although the emphasis is on work done during the International Geophysical Year, particularly that performed with rockets, all other aspects of research into the physics of the upper atmosphere by indirect means are fully covered.

The authors, in their attempt to make their complicated and advanced subject matter comprehensible to a large number of non-specialist readers, have avoided mathematics as far as possible and started with an introductory chapter entitled ' The relevant physics.' This deals briefly with basic ideas used in later chapters relating to electricity and magnetism, wave motion and atomic and molecular processes. It is doubtful, however, to what extent they have achieved this object as, even with this arrangement, it seems that a non-scientist would have considerable difficulty in understanding some of the later sections. A short chapter summarizing the main features of the atmosphere then follows and the remainder of the book deals with each of the main subjects. These in turn are the use of balloons and rockets as research vehicles; probing with sound waves; probing with radio waves; the ozonosphere and the ionosphere; lights in the night sky; aerial tides and magnetic effects; solar, magnetic and ionospheric disturbances; meteors; cosmic rays; and artificial satellites and their potential applications. The final chapter discusses the future possibilities of present research and the manner in which this can be extended.

The subjects listed above are all discussed in some detail and with great clarity. The basic instrumentation and experimental techniques are outlined, the results so far achieved described and the problems which remain unanswered are listed.

With all these excellent features it seems regrettable, from the point of view of workers in related fields, that the attempt was made to simplify the reading. For instance, in one section, care was taken to avoid the use of calculus in deriving the familiar hydrostatic equation, but since it has been necessary to introduce much more difficult equations or ideas without explanation in other sections, it is doubtful whether this type of simplification is justified. Moreover, although original papers are quoted, no detailed references are given but merely a recommended reading list. The book is remarkable for its speed of production and even contains accounts of work performed within the last year, including details of recently launched satellites. Some small price has been paid in that a certain unevenness of treatment and one or two small mistakes have been noted.

These criticisms are minor, however, and it must be regarded as a very useful addition to the library of any meteorologist or worker in related fields involving a knowledge of the upper atmosphere. The production is excellent with many plates, some in colour, and numerous line diagrams.

R.J.M.

Rockets and satellites (Vol. vi of *Annals of the International Geophysical Year*). Ed. by L. V. Berkner. London (Pergamon Press), 1958. Pp. xx, 508; 238 Figs.; 39 Tables. 160s.

Following a general introduction by the editor, about one-fourth of the book is devoted to a discussion of the I.G.Y. rocket and racoon programmes, and deals in turn with launching and trajectories, and with the nature of the experiments and schedule of operations planned by individual countries. The remainder of the book is devoted to the subject of earth satellites – their launching, orbits, tracking, temperature control and details of the instrumentation used to provide information for each of the I.G.Y. disciplines concerned with the size and shape of the earth, solar activity and the physical state of the upper atmosphere. General information on satellites is followed by details of the U.S.S.R. and U.S.A. satellite programmes. In Annex II is included a U.S.S.R. account of the orbits achieved by the first two of their satellites and of the nature of the measurements made : the analysis of their results was proceeding at the time of publication.

Of particular interest to the meteorologist are those parts of the book describing the methods of measuring the detailed characteristics of the solar spectrum, the radiation balance of the earth and such upper-atmospheric parameters as wind, pressure, temperature, density, water vapour and meteoritic dust. In the ' not-too-far-distant ' future, lies the prospect of obtaining a continuous global picture of cloud and storm systems.

The avowed aim of the book (as with other I.G.Y. annals) is to cover all aspects of the subject in such detail as will be required, either by those making observations in this field or by those who will later use the observations. The method of achieving this aim is to reproduce, ' in almost their complete original form,' a series of scientific papers. The work shows signs of its heterogeneous and perhaps over-hurried compilation in the somewhat repetitive nature of some of its contents : in small part, too, the treatment is more discursive and even semi-philosophical than seems altogether appropriate. The book is, however, of some historical interest and provides a vast amount of information on this most spectacular of all the I.G.Y. activities.

D.H.M.

Everyday meteorology. By A. Austin Miller and M. Parry. London (Hutchinson), 1st Edition, 1958. Pp. 270; 77 Figs.; 6 Tables; 22 Plates. 30s.

During the past few years there have been a number of books written about the weather, about meteorology, about climatology, about all these subjects and topics associated with them, by geographers or students of geography. The value and need for some of these books has been viewed with some scepticism in the minds of many meteorologists who regard this invasion of their profession with mixed feelings of scorn and a little trepidation. The geographer, however, has the knack of talking to the general public in the latter's own language and at his level of understanding. As a result books of this calibre often meet with considerable popular success.

Everyday meteorology is a book such as this. It presents a fascinating account of the subject of meteorology and the ways in which the weather can affect the lives of everyone of us, as individuals, as a community, as a world society. It is, broadly speaking, factually accurate; the reader may well imagine that he occasionally overhears a Meteorological Office prompter somewhere behind the scenes.

All is well just so long as geography does not trespass too far into the domain of the physicist and the mathematician. Unfortunately, the temptation is often too great and we have on p. 146 the usual clichés to attempt to explain the Coriolis force in everyday language. Although the effect of the Coriolis force on a northward or southward moving air-parcel may be visualized in terms of momentum conservation, the deflection of an air-parcel moving towards the east or west can not be so explained, nor understood. It is surprising that the deflection of the Foucault pendulum, which can be so easily seen at the London Science Museum, is not used more often as the perfect working example of the Coriolis force. There are other cases of loose phraseology; the World Meteorological Organization does not ' control ' Meteorological Services (p. 32.) In Fig. 33, tropical maritime air is illustrated as coming from the east. If this is intended there is no mention of this type of air-mass approaching from this direction in Table 3 or Figure 26. The atmosphere does not press down on every ' inch ' of us (p. 145) and air can, of course, move up gradient (p. 147) providing it has sufficient inertial velocity.

Thickness lines are not isotherms of ' mean temperature ' (p. 151), and jet streams are not intense ' thermal ' winds (p. 151). Plates IX and X would be better entitled ' Pre- and post-frontal cloud,' respectively.

On p. 216 it is stated that British forecasters are critical of mean 30-day charts which, they say, can have little real significance. In view of the very valuable work at present being done in this country on mean 30-day charts a wrong impression might be obtained that these charts are of no value. In any case scientific thought neither is, nor should take, a national orientation but should be individually free. There are, undoubtedly, meteorologists in this country who see as much promise in mean charts as do meteorologists in other countries.

Meticulous phraseology is a requirement of science but it may not seem important to the general reader who will find this book most interesting reading. The price charged is reasonable enough, but the printing in the reviewer's copy is of poor quality on one or two pages.

A.H.G.

Wetter- und Meereskunde für Seefahrer. By J. Krauss and W. Stein. Berlin (Springer), 4th Edition, 1958. Pp. vii, 206; 104 Figs. (including 2 in back cover); 10 Tables; 32 Cloud pictures (apparently water colour) in back cover. DM 19.50.

The first edition of this book (by Krauss) was published in 1917 under the title *Grundzüge der maritimen Meteorologie und Ozeanographie.* The second edition (in 1931) was by Krauss and Meldau, the names which still appear on the cover. The third edition appeared in 1952 under the present title and by the present authors and the new (4th) edition is essentially the same as the third except where recent developments of the subject required attention : for instance, on p. 152, the notice of the availability of weather forecasts and warnings from the German Fishing Protection vessels in arctic waters appears fully up to date. The change of title does not indicate an excursion in popular science, such as might be suggested by the title, ' *Weather for seafarers,*' in English. In German the simple, down-to-earth word is sometimes preferred to the international word even in the titles of scientific institutes. The same preference for familiar expressions runs through the text; the equivalent technical terms in international usage are also given and the book will introduce the beginner to these. This treatment seems appropriate in a serious elementary teaching manual which has already proved its usefulness in the *Seefahrtschulen.* It is also a simple handbook for reference by mariners.

The book constantly emphasizes how greatly sea traffic depends upon the fund of knowledge of weather, ocean currents and ice, built up over generations by contributions from innumerable vessels of all sizes and types, carefully logging and reporting what they observed. The safety and comfort of every sea-going craft, even today, depends upon continuing this tradition and upon the ability in situations of danger or isolation to interpret what is observed. The plan of the book conforms with this philosophy. First, instruments, measurements and observing methods are dealt with, next, keeping the observation register. After this the basic laws of meteorology and the facts of world climatology and oceanography are simply – perhaps too simply – presented. About twenty pages are devoted to tropical storms, their recognition and manoeuvres to be adopted to avoid the worst of the storm. Another chapter contains a cleverly devised guide to weather advice by radio, including an introduction to codes which gives the hang of them (and thus inspires the beginner's confidence) in a readable manner by means of examples but without full details. The reader is referred to the publications of the *Seewetteramt Hamburg* and of the *Deutsches Hydrographisches Institut* for details and for more advanced theoretical ideas.

The merits of the work include specific examples of weather processes well described (e.g. the northers of the Gulf of Mexico, the *pamperos* of the River Plate, the diurnal fogs of the Californian coast transported by the sea breeze and the easterly swell from the Straits of Gibraltar noticeable as far as 12°W). Also numerous important practical tips, such as ' if you have been prevented from doing an observation at the regular hour, do not attempt to compile an entry in the register from memory. It is better to leave a blank.' The meteorological tips are also well put, e.g., In home waters, if the barometer fails to rise much after a cold front and the wind backs again, the worst storm is still to come in a post-frontal trough.

There are, however, faults, especially of over-categorical statement, to which elementary text-books are prone. For instance (p. 31) ' if the ice is more than 15 cm thick, the ship must have the help of an ice breaker ' and (p. 99) ' depressions . . . . in families, in which the next depression always goes somewhat farther south than the one before.' There is some clumsy wording, e.g. (p. 53) ' the decrease of air temperature to 10 km and constancy of temperature from this level up indicate an inversion at the 10 km level.'

There are various other inaccuracies and mistakes of emphasis, in some cases attributable to insufficient amendment of earlier editions, – e.g., (p. 65). The 500 mb level is equated with cirrus level in our latitudes. The account of the mistral (pp. 79-80) is good, including the curious preservation over the open sea as far as Sardinia of the sharp boundaries of the fast windstream funnelled through the Rhone Valley. Yet the account of local winds is marred at the outset (p. 77) by the lumping together as *Fallwinde* of such unlike phenomena as the Bora, Foehn and Scirocco. The Scirocco (p. 80) is treated as a hot, dry south wind of the western Mediterranean, warmed by Foehn effect over the Atlas : all this is true near the Algerian coast, but the name is Italian and it is in the central Mediterranean that the wind is most frequent – its prominent characteristics there are all due to the excessive moisture content which it quickly acquires over the sea. The reviewer is no oceanographer but suspects that the account (p. 132) of the Gulf Stream as fundamentally a density current is not generally accepted.

The main appeal of this book will be in German-speaking countries, though it contains some points of broader interest regarding the history and current development of the weather service for seafarers provided from Germany. For instance, one learns that thermometer screens on board ship have been abandoned in favour of chromium-shielded sling thermometers. There is also a note of the forecasts in English and radio-picture (*facsimile*) transmission of weather maps to the Arctic fisheries.

H.H.L.

Combination of observations. By W. M. Smart. Cambridge (University Press), 1958. Pp. xiv, 253; 33 Figs.; 25 Tables; 3 App. Tables. 35s.

Here is a new book by the author of that most interesting work, *The origin of the earth*, and of numerous other lucidly written works on astronomy and navigation. The present text-book is arranged clearly and expressed simply.

Details of computing are adequately given for such things as the mean (direct or weighted), moments, correlation coefficient and the fitting of linear equations; but the book is concerned mainly with the theory behind such operations and, throughout, the treatment is alternatively rigorous or completely honest – as occasion demands. The more elementary topics discussed include frequency distributions and associated measures; types of error; the normal law; standard errors; weighting; and simple correlation and regression. In addition some 50 pages are devoted to equations of condition and the solution of the normal equations arising in a least-squares fit of linear relations; Pearson's curves are the principal topic in a longish chapter on theoretical frequency distributions where, unfortunately, the only data used for numerical illustration seem *not* to be frequencies; and a theoretical treatment of normal bivariate and trivariate distributions occupies part of the chapter on correlation. A chapter on ' correction ' of statistics also is best left for a second reading.

Much of the subject matter and a similar treatment appear in Brunt's book (2nd edition 1931) of the same title and publisher; but Smart delves more deeply into mathematical theory, with continuous frequency distributions, Beta functions, Hermite's polynomials and the like. It may well be that the title is a contraction of *Theoria combinationis observationum erroribus minimus obnoxia* given by that other astronomer Gauss to a memoir of his, almost 150 years ago; but both Brunt and Smart, on this subject, are enriched far beyond the scope of Gauss, largely by the enormous contributions to statistics of Karl Pearson (1857-1936).

The new book, with its simple explanation of some methods of statistics and its theoretical discourse on frequency distributions, will be of interest to any worker whose raw material is observations. Logical arrangement and clearness of print and diagrams also make the volume easy for reference.

N.G.

Obituary Notice

Sir GILBERT WALKER, C.S.I., F.R.S.

Sir Gilbert Thomas Walker, C.S.I., Sc.D., F.R.S., Fellow of our Society since 18 January 1905, President 1926-7, Symons Gold Medallist 1934, and an Honorary Member from 1952, died on 4 November 1958 at the age of 90.

Walker was educated at St. Paul's School and at Trinity College, Cambridge, to which he gained a mathematical scholarship. He was Senior Wrangler in 1889 and was elected to a Fellowship of his College in 1891, becoming a university lecturer in mathematics in 1895. Walker's early interest was in electrodynamics, on which he wrote a number of papers from 1892, the winning Adams Prize essay in 1900 being among them. For this and work on a number of dynamical problems such as spinning tops and projectiles he was elected to the Royal Society in 1904. He wrote a treatise entitled ' The theory of electromagnetism ' which was published in 1908.

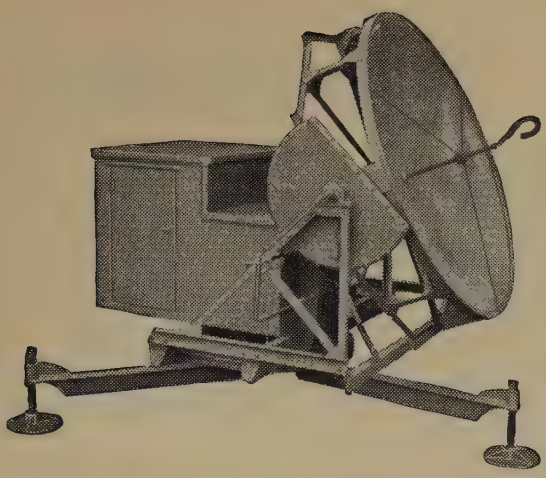
Walker went to India in 1904 to become Director-General of Observatories in succession to Sir John Eliot and he remained in this post until he reached retiring age in 1924. He evidently encouraged his staff to follow their bent in whatever meteorological direction it led, and two of them who joined him from this country, Sir George Simpson and Sir Charles Normand, both past-presidents of this Society, made a big impression on their fields of atmospheric electricity and meteorological thermodynamics respectively. Walker himself, however, became engrossed in the problem of monsoon forecasting which is of such vital concern to much of India, particularly in respect of drought or famine years. Lacking any quantitative theory of the monsoon and therefore of deviations from the normal, Walker frankly followed empirical statistical methods, seeking sizable lag correlations between certain meteorological events in and outside India and the Indian monsoon itself. The danger of the method is obvious : that a high correlation between two events over a series of years is, on its own, no guarantee that the correlation will continue into the future and so provide a sound basis for forecasting. Any two series of random numbers will, in general, over some part of their course, display a high correlation. Walker was well aware of this danger and sought by the most exacting methods to test the significance of his results. It is not possible here to enter into the success of the methods adopted by Walker for monsoon forecasting – it has been fully discussed by Normand in his presidential address to this Society in 1953. The work led Walker to a statistical examination of the relation between events, in many parts of the world, which might or might not be expected to show physical connection, a notable case being the Southern Oscillation – the removal of air from the Pacific Basin between Tokyo and South America and the simultaneous accumulation of air in and around the Indian Ocean, and vice versa. This work led to a long series of memoirs to this Society in the 1920's and 1930's under the title of ' World Weather.' Walker's hope was presumably not only to unearth relations useful for forecasting but to discover sufficient and sufficiently important relations to provide a productive starting point for a theory of world weather. It hardly appears to be working out like that. We still do not have a quantitative theory of the general circulation but such progress as has been made in this study since World War II derives from different beginnings. Nevertheless, as theory advances, it may well be profitable to look back and see whether Walker's findings give a lead for further advances.

Walker, on retirement from India, succeeded Sir Napier Shaw as Professor of Meteorology at Imperial College, University of London. There he not only continued his research on world weather, but took up, with a number of students, the experimental study of convection in unstable fluids with or without shear. These experiments were mainly carried out at modest Rayleigh numbers so that the more vigorous kinds of convection were excluded. But within their range they demonstrated the principles underlying the form of many clouds involving slight instability in the atmosphere. It is possible that this work was stimulated by Walker's earlier fascination, in India, by the soaring flight of birds. The work at Imperial College was certainly attended by a strongly developed interest in and support of the British gliding movement.

Walker retired to Cambridge from Imperial College in 1934 and continued to produce occasional papers on the subject of world weather, and the critical use of correlation in such studies, until 1947. He was editor of our *Quarterly Journal* from 1935 to 1941.

No account of Walker's professional career can begin to give the full flavour of the man. His interests were almost unbounded but highly disciplined. From an early visit to Australia he was led to investigate the design and dynamics of the boomerang. In 1897 he contributed a paper to *Phil. Trans. R.S.* (Series A), and, in 1901, wrote a paper in the *Physikalische Zeitschrift* on the subject. Working models of his own design and construction are still the prized possession of a few. Then, a little later, he wrote a detailed and scholarly article on ' Spiel and Sport ' in the *Encylopädedie des Mathematische Wissenschaften*. Music was also a great interest to him, and he became an acknowledged expert on the theory and evolution of the flute, some modern forms of which are made to a Walker design. How well the writer remembers his last meeting with Sir Gilbert who, already well into the eighties, was on his way to the Royal College of Music for a lesson on the flute. He was engaged on writing a book on this instrument in the last few years of his life. Withal, Walker was eminently friendly – though somewhat reserved – kind, and a very gentle man. The Society mourns the loss of a most distinguished Fellow who served it more than well.

P.A.S.



The Decca Type WF 1 Windfinding Radar, specified for over 40 installations, is tried and proven in operation from Antarctica to the Equator.

Decca Radar

Full operational training facilities are provided for Decca meteorological radar installations. The radars are backed by a world-wide service organisation set up to maintain ground based radar of every kind and marine radar installations in over 8,000 ships.

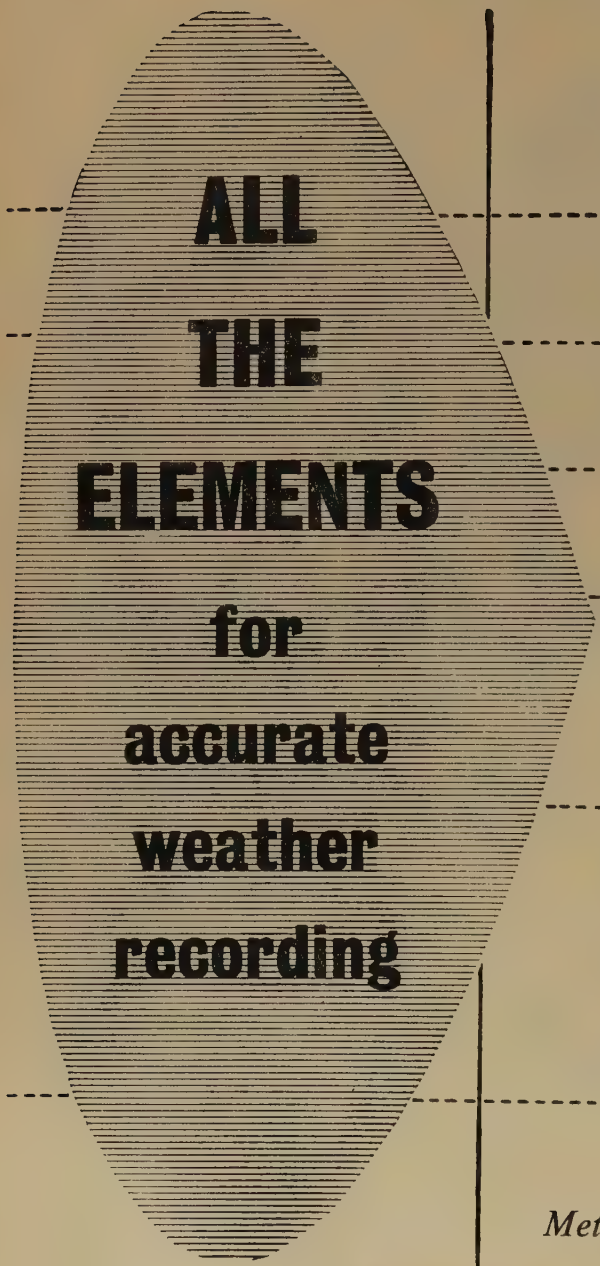
Over 60 Decca Weather radars have been ordered for service in all parts of the world to provide accurate and continuous observation of the position, extent and movement of rain areas.

40 Meteorological Authorities use Decca Windfinding and Weather Radars

DECCA RADAR

DECCA RADAR LIMITED · LONDON · ENGLAND

DR 87



Thermometers

Hygrometers

Barometers

Rain Gauges

Sunshine Recorders

Wind Direction and Velocity Indicators, etc.

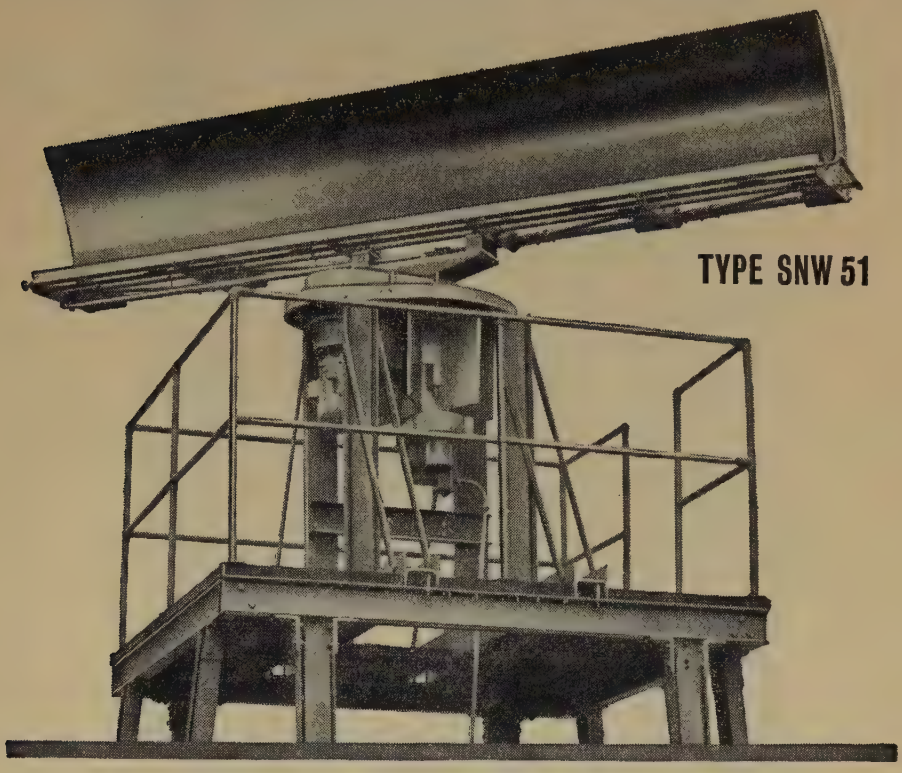
Meteorological Instruments by

NEGRETTI & ZAMBRA LTD

Manufacturers of instruments for the indication, recording, controlling of :— Temperature, pressure, liquid level, volume, specific gravity, humidity, etc.

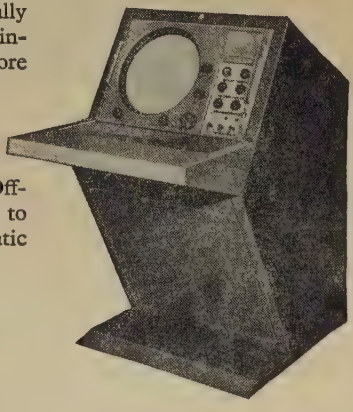
122, REGENT STREET, LONDON, W.1
'Phone: REGent 3406. 'Grams: Negretti, Piccy, London
Branches : Birmingham, Cardiff, Glasgow, Leeds, Manchester, Nottingham.
Agents or Subsidiaries in most countries.

Marconi 3cm Storm Warning Radar



TYPE SNW 51

Marconi's Storm Detection Radar, Type SNW 51 is specially designed to present an accurate picture of storms and rain-producing clouds over ranges up to 200 n.m. It can therefore meet the needs of Meteorological Departments and Airport Authorities, particularly where it is required to derive information from inaccessible areas. Features include:— High Power Working, High-gain Tiltable Aerial (remote-controlled from display unit). Full Off-Centring Facility (enabling any sector to be expanded to cover the whole area of the tube). Remote Displays, Static or Mobile Versions available.



MARCONI

COMPLETE CIVIL AND MILITARY RADAR INSTALLATIONS

MARCONI'S WIRELESS TELEGRAPH COMPANY LIMITED, CHELMSFORD, ESSEX, ENGLAND

M6A

For Your Upper Air Soundings

USE

THE G.I.P. HYDROGEN GENERATOR

Much more economical than aluminium or calcium hydride. Gas production with no sudden increases in pressure. **Used successfully in Adelie Land by the French and in Greenland by the British Expeditions.** Available in different models producing 1, 2 or 3 cubic metres by special patented ready-for-use refills - Excess gas production impossible.

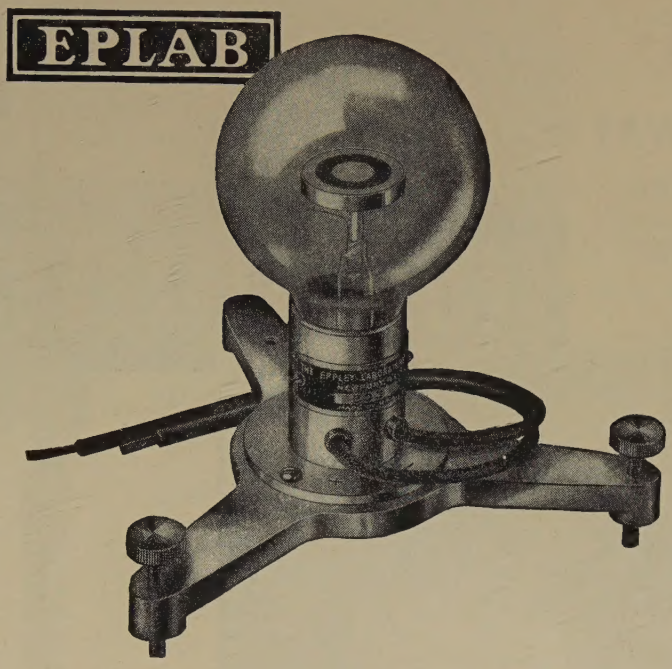
LA G.I.P.

SOCIÉTÉ DES GAZ INDUSTRIELS DE PROVINCE

155, Boulevard Malesherbes, **PARIS.** Téléphone Wagram 22-96

PYRHELIOMETER

For the Measurement of SOLAR RADIATION



Eppley Pyrheliometers are used for solar radiation measurements at ninety-eight weather stations in the continental United States, Canada, Alaska, Greenland, Iceland, Caribbean Sea, and the Pacific Ocean. Sixty-two of these stations are under the direction of the United States Weather Bureau.

The Eppley Pyrheliometer was adopted as standard equipment by the Weather Bureau after considerable experimentation. It was found to be the best instrument so far tested by the Bureau.

Used in conjunction with a suitable recorder, the Eppley Pyrheliometer will provide an accurate and reliable record of total solar and sky radiation on a horizontal surface.

Bulletin No. 2 on Request

THE EPPLEY LABORATORY INC.

Scientific Instruments

20 Sheffield Avenue Newport, Rhode Island, U.S.A.

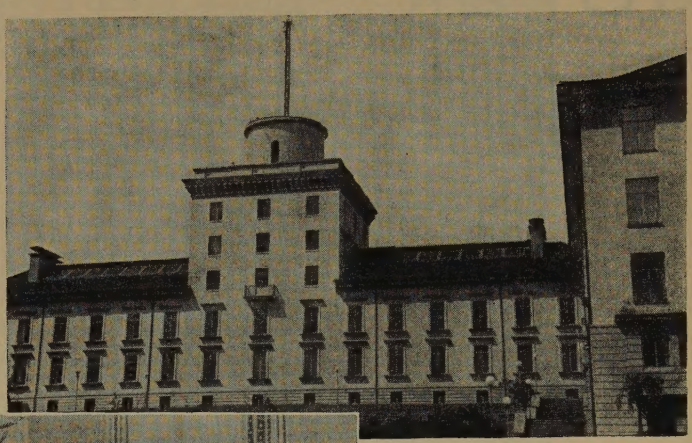
Mufax

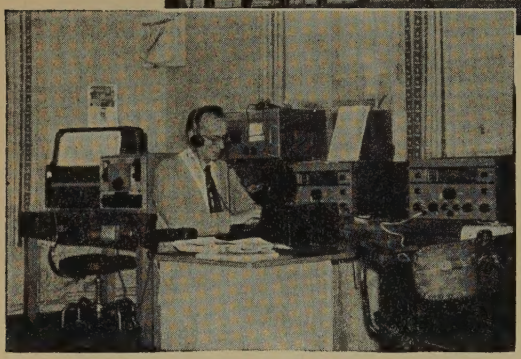
THROUGHOUT THE WORLD

3

NORWAY

BERGEN METEOROLOGICAL INSTITUTE

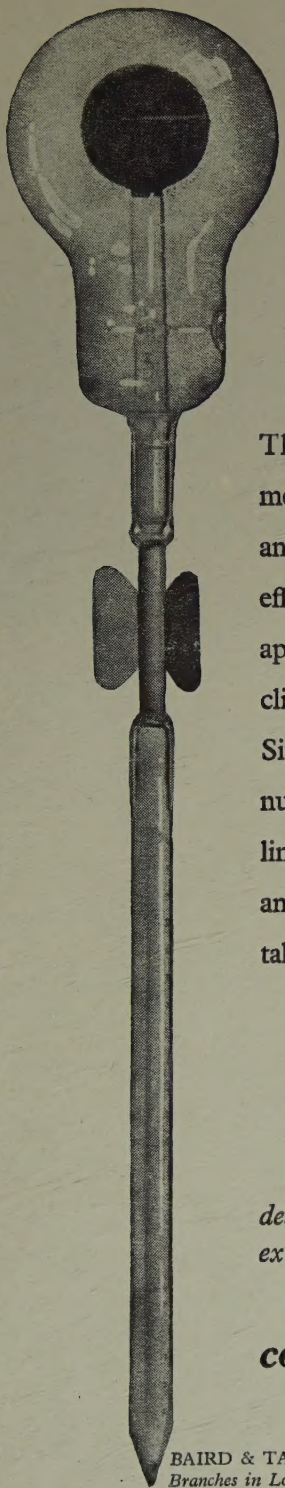




Situated in the tower of the Geofysisk Institutt is the Communications Room of the " Vervarslinga på Vestlandet," the West of Norway Section of the Meteorologisk Institutt. The other floors of the tower are occupied by the remainder of the department under the management of Mr. Finn Spinnangr. The lower photograph shows the radio operator's room with the various receivers, and in the left centre of the picture, a Mufax Chart Recorder D-649.

MUIRHEAD

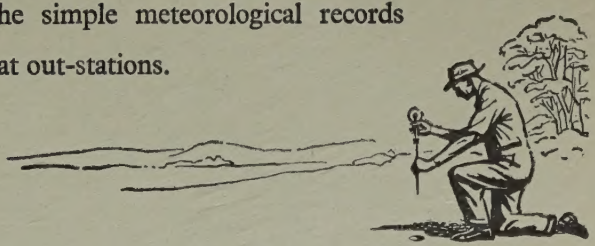
MUIRHEAD & CO., LIMITED,
Beckenham, Kent, England.
MUIRHEAD INSTRUMENTS LTD.,
Stratford, Ontario, Canada.
MUIRHEAD INSTRUMENTS INC.,
441 Lexington Ave., New York 17, N.Y., U.S.A.



Gunn-Bellani radiation integrator

The thermal efficiency of the century-old Bellani instrument has been improved by Gunn, Kirk & Waterhouse, and later by Pereira, from 4% to over 75% thermal efficiency. In its new form, the instrument has extensive applications in meteorology, agriculture, irrigation, micro-climatology, bio-physics etc.

Simple and reasonably priced, it can be used in adequate numbers for field observations, thus providing an essential link between expensive recording thermopiles and the simple meteorological records taken at out-stations.



Write for Technical Leaflet No. T.86, which describes the improvements and gives the results of extensive comparisons with other instruments.

complete laboratory service

BT L

BAIRD & TATLOCK (LONDON) LTD., CHADWELL HEATH, ESSEX, ENGLAND.
Branches in London, Manchester, Glasgow. *Agents throughout U.K. and all over the world.*
TAS/BT.7

F. DARTON & CO. LTD.

Makers of :

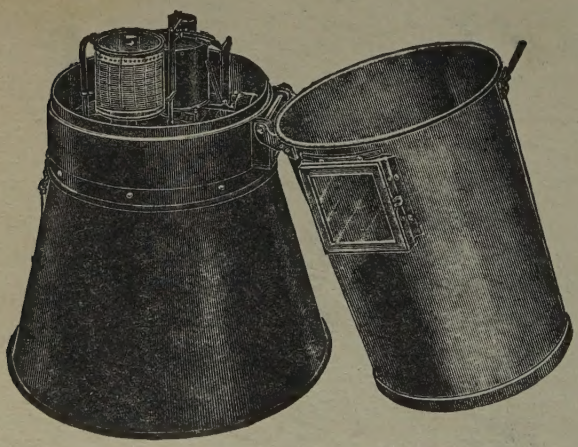
THERMOGRAPHS
BAROGRAPHS
HYGROGRAPHS
PRECISION HAIR HYGROMETERS
ALTIMETER CALIBRATORS
KEW PATTERN BAROMETERS
FORTIN'S PRINCIPLE BAROMETERS
GRASS & SOLAR THERMOMETERS
RAIN GAUGES, etc.

All enquiries to : Established 1834

WATFORD FIELDS, WATFORD, HERTS

Tel. : WATFORD 6019 & 6010

THE DINES
TILTING SYPHON RAIN GAUGE



This is the very latest type of Self-recording Rain Gauge, fitted with a tilting arrangement which ensures a complete syphoning when the rain water has filled the float chamber.

The Instrument Measures
Overall height – 34″
Diameter of base – 20″

Price and further particulars on application

Manufactured and supplied by

HEATH, HICKS & PERKEN (THERMOMETERS) LTD.

NEW ELTHAM, LONDON, S.E.9

Phone : ELTham 3836

Grams : "Optimus, Souphone," London

Printed in Great Britain in Horley Series by The Salisbury Press Ltd., Publicity House, Wilton Road, Salisbury, Wilts. (14283)